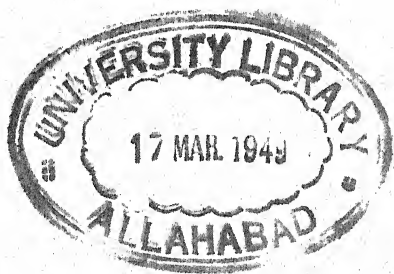


FOREST SOILS



FOREST SOILS

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PREFACE

Silviculture and agriculture are both concerned with the sustained production of plant crops from the soil, which is the land. The silviculturist, like the agriculturist, now recognizes that soil is a basic resource, in the most literal sense of the term, influencing the crops that may be grown, the yields that may be obtained, and even some of the technical details of culture. There are, however, fundamental differences between forestry and agriculture which make a common approach to the problems of soil management in these two fields impractical.

This book was written with two objectives in mind. The first objective was to provide a textbook covering the fundamentals of soil science with applications of the science to forestry which would be suitable for use in the course in soils required of forestry students in American colleges and universities. The second objective was to provide a source of reference to reports of the more important researches on forest soils. This aim necessitated the inclusion of numerous references to foreign as well as domestic literature. The authors realize that the citations of original sources may not meet with the approval of all readers, but they hold the opinion that omission of these references would seriously detract from the scientific value and general usefulness of the book. It is their belief that the thoughtful student and the research worker are interested in the original literature of the field and that they will find the references helpful. On the other hand, readers who do not have this interest should not find the citations objectionable.

Inclusion of a thorough treatment of the fundamentals of soil science has seemed desirable. It is the experience of the authors that standard textbooks on soil science which were designed for agricultural students are not satisfactory in courses for foresters. Use of these textbooks results in the forestry student covering considerable material that does not apply to forest soils or that applies more directly to agricultural soils. As a consequence there

is a reduction in the amount of time that can be devoted to the forestry applications. The student of forestry is as little concerned with tillage practices applicable to agricultural soils as is the student of agriculture with the characteristics of the humus layers of forest soils.

Because of the broad scope of the field covered and the usual limitations of space in a textbook designed for use in a one-semester course, it has been impossible for the authors to treat fully all aspects of forest soils. It is suggested that the instructor will find it a simple matter to supplement this textbook by lectures and assigned reading from the original sources whenever special interests or local conditions make exhaustive treatment of a certain subject desirable. Technical terms peculiar to soil science are usually, although not invariably, defined or explained the first time they are employed. For definitions of technical terms whose meanings are not clear the student should be encouraged to consult standard glossaries, such as A Glossary of Special Terms Used in the Soils Yearbook, in *Soils and Men*, Yearbook of Agriculture, U. S. Department of Agriculture, 1938, and *Forestry Terminology*, compiled and edited by the Committee on Forest Terminology, Society of American Foresters, 1944.

In spite of the progress that has been made soil science still is a relatively youthful discipline. This fact accounts for the tentative nature of available knowledge on certain aspects of soils and makes dogmatic statements and broad generalizations more than usually hazardous. In presenting the subject the writers have sought to avoid both the Scylla of dogmatism and the Charybdis of indetermination.

It is hoped that this treatment of the fundamentals of soil science and interpretation of the more important literature on forest soils will be found useful by foresters, soil scientists, ecologists, and conservationists. To the end that its usefulness to these groups may be increased, the authors invite suggestions for improvement.

New Haven, Connecticut
Ithaca, New York

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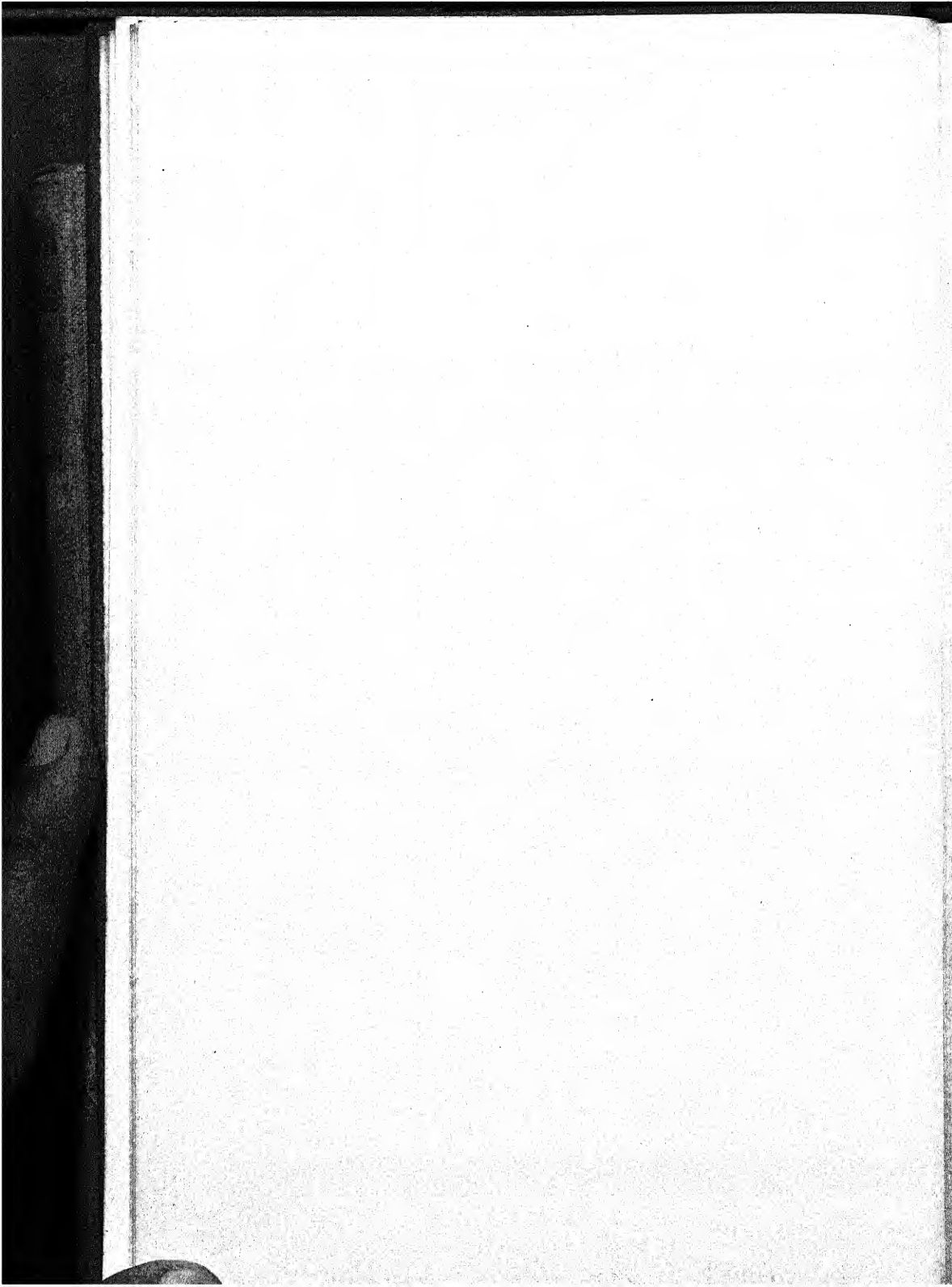
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Chapter 1

INTRODUCTORY

A soil is more than a mere mixture of weathered rock and decayed organic matter. Half a century ago Shaler, an eminent American geologist, recognized this fact and emphasized that the student of soils

... should clearly see that this mass of débris, which at first sight seems a mere rude mingling of unrelated materials, is in truth a well-organized part of nature, which has beautifully varied and adjusted its functions with the forces which operate upon it. Although it is the realm of mediation between the inorganic and the organic kingdom, it is by the variety of its functions more nearly akin to the vital than to the life-less part of the earth. It is not unreasonable to compare its operations to those of the plants which it sustains, for in both there are the harmonious functions which lead matter from its primitive condition to the higher estate of organic existence.

Marbut (1935) defined soil as follows:

The soil consists of the outer layer of the earth's crust, usually unconsolidated, ranging in thickness from a mere film to a maximum of somewhat more than 10 feet, which differs from the material beneath it, also usually unconsolidated, in color, structure, physical constitution, chemical composition, biological characteristics, probably in chemical processes, in reaction, and in morphology.

This definition distinguishes between soil and parent material and emphasizes the characteristics of the soil as a natural body. Soil may also be defined as "the natural medium for the growth of land plants on the surface of the earth" or as "a natural body on the surface of the earth in which plants grow, composed of organic and mineral materials." These two definitions, which are in current use by soil scientists, have the advantage of being edaphological, in that they focus attention on the intimate relationship between the soil and plants.

Among the features which differentiate the soil from other parts of the earth's crust the following may be mentioned: (1) certain

well-defined groups of soils are confined to distinctive climatic regions of the earth, whereas geological formations are not; (2) soils generally are layered, the morphology of the body depending on the conditions under which it developed; and (3) the soil bears a close relationship to the plant and animal life of the earth, the influences being reciprocal.

Recent years have seen a marked impetus in the investigation of the soil as a specific phase of natural philosophy. In other words, interest has been aroused in the scientific study of the soil as a naturally occurring body, comparable to study of minerals or rocks or other natural objects, entirely apart from economic considerations. The term *pedology* has been proposed for this scientific study and has gained wide acceptance. No less scientific, and certainly no less important, are investigations of the soil in relation to the production of plant crops. The term *edaphology* has been used to designate study of the soil from this point of view. Some workers prefer the expression *soil science* to either of the other terms.

DEVELOPMENT OF SOIL SCIENCE

The beginnings of soil science may be traced back to the ancients (Jarilow, 1913). Glinka (1927) reported that soil descriptions were already in existence in the seventh century before Christ. With the gradual development of scientific thought it was natural that attention should be increasingly devoted to the soil. Scientists in many different fields contributed to the development of soil science; geologists, in particular, have rendered exceedingly valuable service. Among the early investigators who made important contributions were Berzelius in Sweden, Ebermayer, Fallou, Friedrich Karl Hartig, Heyer, Hundeshagen, von Liebig, Ramann, Richthofen, Sprengel, Thaer, and Wollny in Germany, Dokuchaiev, Gedroiz, Glinka, and Sibirtzev in Russia, Davy, Darwin, Lawes, and Way in England, Müller in Denmark, and Hilgard, Hopkins, Johnson, King, Marbut, Ruffin, Shaler, and Whitney in America. References to the work of many of these investigators will be made in the following chapters.

Attention was first directed to agricultural rather than forest soils. This was natural, because agricultural problems arose

earlier and were more pressing than those of forestry. In certain European countries, however, as the practice of forestry developed, the need for information on forest soils became more and more apparent. In 1830 Hundeshagen published his 280-page book, *Die Bodenkunde in land- u. forstwirtschaftlicher Beziehung*, which was followed in 1856 by Heyer's *Lehrbuch der forstlichen Bodenkunde und Klimatologie* and in 1876 by Ebermayer's *Die gesammte Lehre der Waldstreu mit Rücksicht auf die chemische Statik des Waldbaues*. The German edition of Müller's classical *Studien über natürlichen Humusformen und deren Einwirkung auf Vegetation und Boden* appeared in 1887, and Ramann's *Forstliche Bodenkunde und Standortslehre* became available in 1893. These books were among the earliest which dealt rather specifically with forest soils, as contrasted to agricultural soils. Many of the concepts set forth by these pioneer writers are still regarded as sound.

In America active investigation of forest soils is a recent development, extending back over a period of only fifteen or twenty years. As might be expected, therefore, American literature on forest soils is limited; information on many questions is almost wholly lacking. Fortunately, it is unnecessary to rely entirely on information derived from American sources; much of value is found in the accumulated facts and experience of investigators and practitioners in other countries. The fundamental principles of soil science and plant physiology are equally valid in all parts of the world. It should be pointed out, however, that caution and judgment must be exercised in applying foreign practice to American forest conditions. In no other field, perhaps, does broad generalization involve greater dangers than in that of forest soils.

FOREST CROPS AND AGRICULTURAL CROPS

In the study of forest soils much may be gained from the experience and information of workers in agricultural soils. However, agricultural crops differ from forest crops in many respects; cultural operations of the agriculturist and the forester stand in decided contrast. Among others, Burger (1926: 82) has emphasized the necessity of caution in applying the experience of the agriculturist to the forest. Many investiga-

tors regard the physical properties of forest soils as of outstanding importance, whereas in agricultural soils the chemical properties may be most important (Burger, 1924; Leiningen-Westerburg, 1925: 136; Krauss and Härtel, 1935: 214). In general, the requirements of forest trees for soil nutrients are much less than those of most agricultural crops (Ebermayer, 1876; Morosov, 1928).

The plants grown by the agriculturist are generally of either exotic or domesticated varieties, whereas the forester is usually concerned with species represented in the natural vegetation of his region. In agricultural production the plants grown may not belong naturally to the soil, so that artificial conditions must be created to insure their development. Furthermore, in agriculture there frequently exists a desire to stimulate extraordinary growth in localized organs of the crop plants; such stimulation usually requires abnormal soil conditions, particularly from the standpoint of nutrients (Fraser, 1922).

A further difference between agricultural and forest crops is the length of time required for maturity. The agriculturist usually harvests each year the whole plant, exclusive of the roots; to compensate for the large amount of nutrient materials removed he fertilizes the soil. In forestry many years generally elapse between the establishment of the crop and the harvest. Meanwhile, substantial quantities of organic debris are returned to the soil each year in the leaf fall. This organic debris exerts an important influence on the physical, chemical, and biological characteristics of the soil. The mineral content of the wood removed from the forest at the time of harvest is relatively small. Forest vegetation has a more profound influence upon the soil climate than does an agricultural crop. It has been pointed out by Dengler (1935: 104) that the presence of forest vegetation results in what amounts to an assumption by the soil climate of a more oceanic nature. Essentially the same view was advanced by Albert (1928), who regarded agricultural use of land as tending to create a steppe soil climate. From the standpoint of temperature relations forest vegetation creates conditions which amount to a shift of the soil climate 1° - 2° latitude to the north.

Another marked difference between forest crops and agricultural crops is the generally periodic financial returns from the forest crops, as contrasted to the annual returns from agricultural

crops. In addition, returns from forest crops are commonly lower than from agricultural crops.

In agriculture, because of the generally higher value of the crop per unit area, much larger expenditures for cultural operations can be justified than in forestry. This fact means that the soil can be cultivated and that fertilizers can be supplied as they are needed. In forestry, cultivation of the soil is generally not feasible, or at best it is carried on only as a preliminary to the establishment of a new stand; even then it is relatively simple and inexpensive. Fertilization of forest soils likewise is generally not feasible. Lang (1920) has pointed out that mistakes in the management of a forest soil do not become apparent as quickly as in agricultural soils. Such mistakes, however, cannot be quickly corrected.

Maintenance of fertility is a highly important problem in the management of both agricultural and forest soils. It is fundamental that in humid regions nutrient materials are moved downward in the soil body by the descending water. Materials carried downward from the surface layers may be (1) held by chemical or physical forces in the deeper soil layers, (2) absorbed by plant roots and returned to the soil surface in the leaf fall, or (3) permanently removed from the soil body in the drainage water. The many techniques applied to agricultural soils for the maintenance of fertility include rotation of crops, fertilization, tillage, and the prevention of undue erosion. The forester, however, must resort to other means. In general, it may be said that forest-soil fertility can best be maintained by the development of healthy, full stands of species adapted to each particular site. Under these conditions the soil body is permeated to a considerable depth by active root systems, which Walter (1936) has likened to a filter. The high importance of full utilization of the soil body by plant roots as a means of maintaining fertility has also been recognized by Albert (1928: 117). Nutrient constituents in their downward movement through the soil, together with materials originating in the deeper layers, are absorbed by the roots, translocated to the above-ground organs of the trees, and in due time returned to the soil in the leaf fall. In this way the nutrient capital of a forest soil may be largely concentrated in the upper horizons, where most of the roots occur. In good forest soils, where decomposition of organic debris is fairly rapid, there

is a rather rapid turnover of the nutrient capital of the soil, the nutrient materials being used over and over again. At any given time the total nutrient capital of a site is represented by the soil and the vegetation it supports. By plowing, the agriculturist is to some extent able to return to the surface materials which have been translocated downward; the forester, however, must depend on absorption by tree roots, activity of the soil fauna, and natural agencies, such as windthrow.

FOREST SOILS AND AGRICULTURAL SOILS

Forest soils differ from agricultural soils in many respects. Certain of these differences are inherent and may be traced back to dissimilarities that existed even in the virgin condition. In general, the soils in agricultural use were selected because of certain physical and chemical properties which rendered them desirable.

In most parts of America the forest soils are those which are relatively undesirable for agriculture. The better soils have been cultivated; the poorer have remained in native vegetation. The soil in many forest areas is exceedingly poor, so poor, in fact, that it cannot be used for the production of agricultural crops. On the other hand, highly productive soils also may be devoted permanently to forest production because certain qualities, such as the presence of rocks, render them unsuited for cultivation. Forest soils commonly contain more large rocks and may be less deep than agricultural soils. Although the influence of rocks in the soil body is very imperfectly understood, there is no reason to doubt that this influence exists. As a result of selection it is fair to assume that in any given region the agricultural soils do not exhibit the degree of variability found in the forest. Frequently forest soils are younger than agricultural soils, and their chemical constitution may be unbalanced or deficient when they are derived from a single geological formation (Hilgard, 1892: 13; Bates, 1928). For this reason, as will be emphasized later, the nature of the parent material is of particular importance in forest soils.

A forest soil, according to Romell (1930), is a natural or a slightly modified natural product, whereas an agricultural soil is an artificial product. Forest soils in the virgin condition gen-

erally exhibit a well-defined succession of natural horizons. Of particular interest and importance are the organic layers. In most agricultural soils the purely organic layers are lacking, and the upper horizons are profoundly disturbed by cultivation, with the result that an artificial horizon is developed in the upper six inches or so of soil. Krauss and Härtel (1935: 212) regarded cultivation, which has both vertical and horizontal influences, as favoring greater uniformity in agricultural soils. Certainly this is true in the soil zone subjected to intimate mixing, that is, the zone of cultivation. The concept of soil as a natural body should appeal more to the forester, possibly, than to the agronomist; the forester usually studies soils in their virgin or nearly virgin state, whereas the agronomist seldom deals with undisturbed soils. Differences between these two groups of soils are enhanced by the use of fertilizers. In agricultural soils various nutrient materials are commonly added. The effect of these fertilizer materials is to create artificial chemical conditions in the soil.

SOIL AS A PRODUCT OF ENVIRONMENT

A soil may be regarded as a product of its environment. The influences which affect the development of the soil body are many; in fact, so great is their number that Mückenhausen (1936) has appropriately referred to them as the "constellation" of soil-forming factors. A soil is not a static body but is dynamic and may be expected to change with modification of its environment. The factors, or factor groups, most important in determining the character of the soil may be indicated as *climate, living organisms, relief, time, and parent material*. All these factors merit consideration, and their influences will be discussed in later sections. At this time particular attention is called to the role of vegetation in soil development. Braun-Blanquet (1934) has pointed out that in central Europe there is a close relation between the natural evolution of the vegetation and the evolution of the soil. It also appears, in some cases at least, that quality of the soil is in large measure a function of the forest stand it supports (Burger, 1926: 80). Blanckmeister (1938) has expressed the opinion that the influence of a specific forest stand on the soil is greatest in regions climatically favorable for the development

of that vegetation type. Vegetation influences the soil, and, of course, the nature of the soil also influences the vegetation.

GENERAL RELATIONS OF PLANTS TO SOIL

The soil is the natural medium for the growth of land plants; it provides a foothold, and from it are obtained the necessary supplies of water and nutrients. At first thought it may seem relatively simple to specify the plant-producing capacity of a soil in terms of its physical and chemical characteristics. Such specification, however, is not easy. Süchting (1930: 587) and Wiedemann (1934) have emphasized the fact that soil productivity is determined by the *integrated effect of all its characters*, rather than by the magnitude of any one. The so-called law of compensating factors needs to be mentioned in this connection. The idea embodied in this law is simply that excellence in one factor may compensate, within limits, for poverty in another. A factor which is present in minimum need not be strictly limiting; other factors which are unusually favorable may enable the plant to utilize better this factor in minimum (Hesselman, 1926; Wittich, 1930; Wiedemann, 1934).

The farther a tree species is removed from the region of its climatic optimum, the more discriminating it becomes in respect to soil (Mayr, 1925: 79; Blanckmeister, 1938). Thus the soil factor becomes increasingly important near the limits of the climatic range of a species.

It should be clearly appreciated that soil and habitat are not synonymous; although highly important, soil is but one factor of a habitat. Consequently it is impossible to specify site quality solely in terms of soil. As early as 1886 this concept was set forth by Grebe.

In considerations of the relation of soils to silviculture the upper horizons, particularly the organic layers, merit special attention. Most of the feeding roots of trees are located in the upper parts of the soil body; these are the horizons that can be most readily influenced by silvicultural operations. The forester has little, if any, control over conditions in the deeper horizons. Certain of the soil characteristics which influence plant growth are essentially permanent; others are rather readily changed. Texture, depth, and topographic situation, for example, cannot be modified

by man, whereas such characteristics as structure and content of organic matter are susceptible to change, in either a favorable or an unfavorable direction. Forest-soil factors which may be influenced by stand composition, stand density, grazing, and fire are of particular interest to silviculturists.

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Chapter 2

SOIL-FORMING MINERALS

Minerals, which are naturally occurring substances of definite chemical composition produced by inorganic processes, are usually composed of two or more elements. Rocks generally consist of aggregates of two or more minerals, although some are composed chiefly or entirely of one mineral. The importance of minerals and rocks in soil science is obvious from the fact that the solid inorganic phase of soils consists of mineral fragments and their alteration products.

More than one thousand minerals are known to science, but a very few of them account for over 99 per cent of the weight of igneous rocks (Clarke and Washington, 1924). The number of minerals constituting any considerable portion of a given rock usually is not more than three or four. Estimates by Clarke (1924: 33) of the average mineral composition of igneous rocks, shales, and sandstones are presented in Table 1.

TABLE 1. AVERAGE COMPOSITION OF IGNEOUS AND SEDIMENTARY ROCKS

(After Clarke, 1924)

	Igneous Rock, per cent	Shale, per cent	Sandstone, per cent
Feldspar	59.5	30.0	11.5
Amphiboles and pyroxenes	16.8
Quartz	12.0	22.3	66.8
Mica	3.8
Clay	25.0	6.6
Limonite	5.6	1.8
Carbonates	5.7	11.1
Other minerals	7.9	11.4	2.2
	100.0	100.0	100.0

It will be shown later that these data can be applied only for generalization, since some igneous rocks, for example, contain no

quartz, whereas others have it in great abundance. Likewise some sandstones are very high in feldspar; others are low in this mineral.

Stated as oxides, the most abundant constituents of igneous rocks are silica (SiO_2), alumina (Al_2O_3), ferric oxide (Fe_2O_3), ferrous oxide (FeO), magnesia (MgO), lime (CaO), soda (Na_2O), potash (K_2O), and water (H_2O). These oxides make up 98 per cent of the matter of igneous rocks. Consequently, it is not surprising that silicates and aluminosilicates of iron, magnesium, calcium, sodium and potassium are of great importance in any consideration of rocks and soils.

The rock- and soil-forming minerals may be divided into two groups: (1) original or primary constituents, and (2) secondary constituents. The secondary minerals are alteration or decomposition products of primary minerals. The line between the two groups is not distinct, since certain minerals, for example, quartz and calcite, may be either primary or secondary. The feldspars are primary minerals but on weathering give rise to secondary minerals. The primary mineral magnetite may alter by oxidation to hematite or limonite or, by the action of carbonated water, to siderite.

In the following discussion only minerals that are rock- or soil-forming or are important from the standpoint of plant nutrition are considered. No attempt has been made to include a complete mineralogical description of each individual mineral; for this information the standard works on mineralogy should be consulted.

SOME PRINCIPLES OF THE BASIC STRUCTURE OF MINERALS

Because of recent advances in our knowledge of the atomic structure of minerals (Bragg, 1937; Pauling, 1939) and because of the extreme importance of these principles in understanding the nature of the clay minerals, which will be discussed fully in Chapter 7, it seems advisable to consider briefly in simple terms the arrangement of individual ions in minerals.

The basic consideration is the size of the atoms concerned. Because of the small dimensions involved, all measurements are expressed in angstrom units (one angstrom unit equals one ten-thousandth of a micron and is abbreviated A). An atom, con-

sidered by itself, has no rigid boundary. As Bragg (1937: 30) states, "It is impossible to draw a sphere around the nucleus, which just contains the electrons bound to it. But when two ions approach, a repulsive force which resists any closer approximation sets in with great abruptness near a certain value of the interatomic distance." Thus, atoms in contact can be considered as spheres; each sphere has a radius which is characteristic of the ion involved. It will be noted that both atoms and ions have been mentioned. In reality, the elements in minerals exist as ions; that is, there is a redistribution of electrons among the atoms so that instead of being neutral they acquire positive and negative charges. This concept is important in a consideration of the charge on colloidal clay minerals.

Oxygen is a large ion, having a radius of 1.40 Å. In fact, oxygen is so large that it alone determines the structure of most soil minerals. The common cations, such as sodium, magnesium, iron, aluminum, and silicon, are so much smaller than oxygen that they occur in the interstices between the oxygen ions.

The small cations are surrounded by a group of closely packed oxygen ions arranged in a systematic order. The number of oxygen ions that surround a cation is usually determined by the

TABLE 2. ARRANGEMENT AND MINIMUM RADIUS RATIOS
FOR VARIOUS NUMBERS OF CLOSELY PACKED
OXYGEN IONS

Number of Oxygen Ions	Arrangement	Minimum Radius Ratio
3	Equilateral triangle	0.14
4	Tetrahedron	0.22
6	Octahedron	0.41
8	Cube	0.73
12	Cubo-octahedron	1.00

size of the cation. For example, boron has an ionic radius of 0.20 Å. and usually occurs with three oxygen ions surrounding it, their centers being at the corners of an equilateral triangle. Silicon has an ionic radius of 0.39 Å. and is surrounded by four oxygen ions at the corners of a tetrahedron. Aluminum, with a radius of 0.57 Å., can also occur in the tetrahedral arrangement, or it can occur surrounded by six oxygen ions at the corners of an octahedron. As an expression of this important relationship between the sizes of ions involved, the radius ratio is used. When

large oxygen ions are packed around smaller cations, the radius ratio is obtained by dividing the cation radius by the oxygen ion radius. For example, since oxygen has a radius of 1.40 A. and silicon 0.39 A., it follows that the radius ratio of silicon is $0.39/1.40$, or 0.28. For each number of oxygen atoms that can be closely packed around a cation, there is a minimum radius ratio that can obtain. In Table 2 are shown the possibilities in this respect.

The number of oxygen ions around a cation is referred to as the coordination number of the cation. If the radius of the cation and of the anion is known, the radius ratio can be calculated, and then from the data in Table 2 the coordination number can be predicted. The necessary data and corresponding predictions for some of the ions present in commonly occurring soil minerals are reported in Table 3.

TABLE 3. CRYSTAL IONIC RADII, RADIUS RATIOS, AND PREDICTED OXYGEN COORDINATION NUMBERS FOR CERTAIN CATIONS

Ion	Crystal Ionic Radius	Radius* Ratio	Predicted Coordination Number
Boron ⁺⁺⁺	0.20	0.14	3 or 4
Silicon ⁺⁺⁺⁺	0.39	0.28	4
Aluminum ⁺⁺⁺	0.57	0.41	4 or 6
Iron ⁺⁺	0.60	0.43	6
Magnesium ⁺⁺	0.78	0.56	6
Sodium ⁺	0.98	0.70	8
Calcium ⁺⁺	1.06	0.76	8
Potassium ⁺	1.33	0.95	8 or 12
Barium ⁺⁺	1.43	1.02	12

* These values are based on oxygen as 1.40 A. and on cation radii as given by Pauling (1939) or Goldschmidt (1929). It is assumed that these values may be stated more correctly as further study is accomplished.

The important facts to be learned from Tables 2 and 3 are that silicon occurs only in the tetrahedral position, whereas aluminum may occur in either the tetrahedral or octahedral position. It should also be noted that magnesium and iron occur in the octahedral position. In practice this means that aluminum can replace silicon in the tetrahedral position, and iron and magnesium can replace aluminum in the octahedral po-

sition. Some examples of these replacements will be presented later.

The atomic structure of minerals is a subject for an entire book; only a brief account can be given here. In order to see how fundamental these principles are, however, let us discuss a few practical examples. First, consider the situation in the molten magma of the ancient earth's crust, for example, where an abundance of oxygen and magnesium exists. With plenty of oxygen available, there would be four oxygen ions for every silicon ion; hence silicon-oxygen tetrahedra would develop, with silicon occupying the central position and the four oxygen ions at equal distances from it. Since oxygen has a valence of two and silicon a valence of four, the SiO_4 radical that results would have an extra negative charge of four. In order to balance this negative charge of four, two magnesium ions (an abundance of this element was postulated) could combine with the silicate radical, forming the mineral forsterite, Mg_2SiO_4 . If, instead of magnesium, there was an abundance of zirconium, the mineral zircon, ZrSiO_4 , could be formed; or, if both magnesium and iron were present, the common mineral olivine, $(\text{Mg}, \text{Fe})_2\text{SiO}_4$, might be expected. The long series of minerals in the garnet group are all combinations of cations with this basic SiO_4 group.

The situation, however, is frequently complicated, because there may not be enough oxygen so that every silicon ion can have four oxygen ions, or every aluminum ion can have four or six oxygen ions. When this condition occurs, it is necessary for the tetrahedral or octahedral groups of oxygen ions to share corners or edges or even sides, depending upon the degree of scarcity of oxygen ions. Quartz affords an example of the sharing of corners of tetrahedra, that is, each oxygen ion is shared by two silicon ions. The formula for quartz, therefore, is not SiO_4 , but SiO_2 . Here four negative charges are balanced by four positive charges, and no other cations need be involved.

In order to illustrate the close relationship existing between some of the different minerals, brief consideration will now be given to the feldspars. Four molecules of quartz may be represented by the formula Si_4O_8 . It has been stated that aluminum may be substituted for silicon in the tetrahedral position. If one of the silicon ions is replaced by an aluminum ion, the formula of the resulting compound would be AlSi_3O_8 . This unit is un-

balanced, because silicon has four positive charges and aluminum has only three. Therefore some cation will move in to satisfy the extra negative charge. If the cation is potassium, either orthoclase or microcline with the formula KAlSi_3O_8 will result. If the cation is sodium, the feldspar albite with the formula $\text{NaAlSi}_3\text{O}_8$ will be obtained. If replacement of two silicon ions by aluminum is assumed, there would result two extra negative charges; calcium might move in to satisfy these charges, thus forming the feldspar anorthite, $\text{CaAl}_2\text{Si}_2\text{O}_8$. Actually all sorts of combinations between these extremes are known; part of the negative charge may be satisfied by one cation and part by another. The important thing to note is that there are eight oxygen ions in all these feldspars. In other words, the basic framework of the minerals is made up of oxygen ions in certain definite arrangement. The cations may shift so as to keep the electrical charge balanced so long as the spatial requirements are satisfied. As a point of interest the location of the potassium ion in orthoclase will be mentioned. In Table 3 it may be noted that the potassium ion is nearly as large as the oxygen ion. Therefore it cannot be located in the center of the tetrahedral groups as are silicon and aluminum. What actually happens is that every other tetrahedron is inverted, so that the apical oxygen ion is common to each pair of tetrahedra. This arrangement leaves sufficient extra room for the large potassium ion to occur adjacent to the apical oxygen ion, and it does not interfere with the linked tetrahedral structure of the feldspars.

As yet all the structures discussed have involved only oxygen tetrahedra. However, many minerals have some of their cations at the center of octahedra. They will be discussed in connection with the detailed properties of specific minerals; particularly in Chapter 7 in connection with the clay minerals of soils. The main object of the foregoing discussion is to simplify the concept of mineral structure by showing that it is the result of the closest possible packing of negatively charged oxygen ions around positively charged cations. Substitution takes place among the cations of similar size more or less independently of their nature or valence. When we consider that approximately 95 per cent of the *volume* of the earth's crust consists of oxygen, it is not surprising that this element constitutes the basic structural ma-

terial of the individual minerals. Diagrams of a silicon-oxygen tetrahedron, an aluminum octahedron, and a silicon-oxygen sheet of linked tetrahedra, of the type that occurs in mica and the silicate clays, are shown in Figure 1.

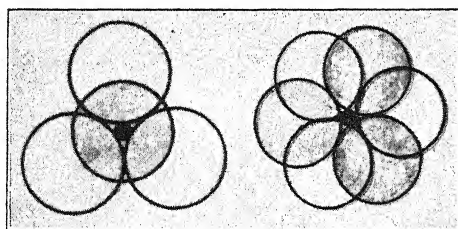
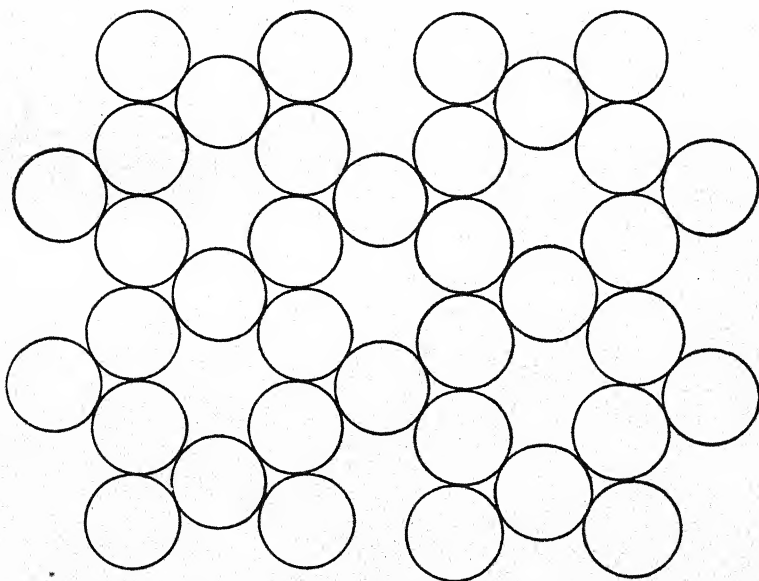
*A**B**C*

FIG. 1. *A*, a silicon-oxygen tetrahedron; *B*, an aluminum-oxygen octahedron; *C*, basal oxygen ions in a silicon-oxygen sheet of linked tetrahedra.

SOME PROPERTIES OF CERTAIN MINERALS

The student of either forest soil science or agricultural soil science is concerned with minerals largely as a direct or indirect source of mineral nutrients for plants. In addition, however, the study of minerals *per se* is interesting, and knowledge concerning them will increase enjoyment while in the field. Therefore the discussion of a specific mineral will emphasize the principal plant nutrients furnished by the mineral, together with some suggestions for identifying the mineral with a hand lens under field conditions. It is impossible to present such identification data adequately in a brief discussion; hence the student is expected to obtain additional guidance from his instructor.

SILICATES

The simple silicates of the alkali metals are not known to occur free in nature, but they form abundant aluminosilicates. Calcium and magnesium, however, form both aluminous and simple silicates. Silicates constitute the most abundant and important class of soil-forming minerals. From them clays are derived; hence they are fundamentally involved in determining the value of soils. In the treatment which follows only the most important silicate minerals are discussed.

FELDSPARS

The feldspars are the most abundant of all minerals, forming nearly 60 per cent of the mineral matter of igneous rocks. All the feldspars may be altered, being rather readily attacked by water containing carbon dioxide; alkaline solutions also bring about their decomposition. The finer particles of soil consist largely of alteration products of the feldspars.

The feldspars are essentially anhydrous aluminosilicates. A portion of the silicon in the oxygen tetrahedra has been replaced by aluminum, and the remaining excess negative charge has been balanced by potassium, sodium, or calcium or by both calcium and sodium.

Microcline, or potash feldspar, has the chemical composition

KAlSi_3O_8 , a hardness¹ of 6.0–6.5, and a specific gravity of 2.56. This mineral is particularly characteristic of the more siliceous plutonic rocks; it also occurs in many eruptives and in metamorphic schists. Disintegration is favored by its cleavage, which is perfect in two planes at right angles to each other. The alteration of microcline to kaolinite by water containing carbon dioxide may be illustrated thus: $2\text{KAlSi}_3\text{O}_8 + 2\text{H}_2\text{O} + \text{CO}_2 \longrightarrow \text{H}_4\text{Al}_2\text{Si}_2\text{O}_9 + 4\text{SiO}_2 + \text{K}_2\text{CO}_3$. The mineral is a large primary source of potassium in soils and probably one of the more important sources of clay minerals. *Orthoclase* is very similar to microcline in composition and occurrence but is monoclinic in form, whereas microcline is triclinic. As a rule these two minerals cannot be distinguished by megascopic examination.

The feldspars as a whole can be identified in hand specimens of rocks, since they are inferior in hardness to quartz and have a very distinct cleavage, which quartz lacks. They may be distinguished from calcite by their greater hardness and their failure to effervesce in acids. It is more difficult, however, to distinguish microcline and orthoclase from the plagioclase feldspars by megascopic examination in the field. Perhaps the most useful criterion is the nature of the twinning² exhibited by the crystals. If on close examination with a good hand lens fresh cleavage surfaces show fine, parallel lines or striations (albite twinning), the conclusion may be drawn that the feldspar belongs in the plagioclase group. Orthoclase, being monoclinic, can not exhibit albite twinning, but microcline, being triclinic, may show twinning according to the albite law. However, this twinning in microcline is not readily detected by megascopic examination. Unfortunately, apparent or actual absence of albite twinning is not in

¹ The scale of hardness commonly adopted is as follows:

1. *Talc*, easily scratched with the thumbnail.
2. *Gypsum*, can be scratched with the thumbnail.
3. *Calcite*, not scratched by thumbnail, but easily cut with a knife.
4. *Fluorite*, can be cut by knife, but less easily than calcite.
5. *Apatite*, can be cut with a knife only with difficulty.
6. *Orthoclase*, can be cut by knife, but only with great difficulty, and on thin edges.
7. *Quartz*, cannot be cut with a knife, will scratch glass.
8. *Topaz*, will scratch glass.
9. *Corundum*, will scratch topaz.
10. *Diamond*, will scratch corundum.

² For a practical consideration of twinning and its identification English (1934: 72–75) should be consulted.

itself proof that a feldspar does not belong in the plagioclase group.

The *plagioclase* (soda lime and lime soda) feldspars embrace a series of minerals consisting of albite and anorthite as the end species, together with several representing isomorphous mixtures of the two. The minerals in this group tend to be more abundant in the less siliceous, that is, the more basic, rocks, such as the gabbros and basalts. Field identification of the various members of the plagioclase group in hand specimens of rocks is generally impossible.

Albite, or soda feldspar, has the chemical composition $\text{NaAlSi}_3\text{O}_8$. Its hardness is 6.0–6.5, as is that of other feldspars, and the specific gravity is 2.60. Albite is commonly found in granites, gneisses, crystalline schists, diorites, and trachytes.

As calcium is substituted for sodium in an increasing degree, the minerals *oligoclase*, *labradorite*, and finally *anorthite*, where complete replacement has resulted, are obtained. Anorthite, called lime feldspar, has the formula $\text{CaAl}_2\text{Si}_2\text{O}_8$. The occurrence of oligoclase is similar to that of albite, but labradorite and anorthite usually occur in basic eruptives, such as the diabases, basalts, and norites. Labradorite and anorthite represent important primary sources of calcium in soils.

ZEOLITES

In the zeolite group belong a rather large number of hydrous aluminosilicates, with varying percentages of calcium, sodium, and potassium. Occasionally they may contain magnesium or barium. This group of minerals does not occur in soils (Robinson, 1936: 82; and Russell, 1937: 65), but they are of particular interest to soil scientists because of their high capacity for holding exchangeable cations, a characteristic in which they resemble clay minerals. They are similar to the feldspars in structure; analcite, for example, has the formula $\text{NaAlSi}_2\text{O}_6 \cdot \text{H}_2\text{O}$. The water is so loosely held that it can be driven off without changing the crystal structure, and it is taken up again upon standing in the air. As in feldspar structure, it can be seen that a silicon ion has been replaced by aluminum, and sodium has neutralized the extra charge. The oxygen tetrahedra are so linked at the corners that rather large spaces are left in the crystal lattice. This arrange-

ment permits the easy passage of cations which exchange with the sodium, thus accounting for the exchange capacity of the mineral. It should not be assumed, however, that the exchange capacity of soils is caused by the presence of zeolites; rather, the mechanism of cation exchange in soil colloids has certain features in common with that in zeolites.

THE PYROXENES AND AMPHIBOLES

These two groups of minerals are closely related and can be considered together. They are members of the so-called ferromagnesian group and are essentially metasilicates of calcium, magnesium, and iron; sometimes they contain aluminum and sodium. The ionic structure of these minerals is complex. In order to show the application of the principles discussed earlier in this chapter it may be stated that the basic framework of the pyroxenes and amphiboles consists of long chains of linked silicon-oxygen tetrahedra, as shown in Figure 2, with the metasilicate formula of SiO_3 . The pyroxenes consist of single chains (Figure 2A), whereas the amphiboles have double chains (Figure 2B). The crystal is built up by linking together two chains in parallel. These chains are held together by the cations; the calcium is surrounded by eight oxygen ions, and the magnesium and iron by six.

In order to identify rock specimens correctly it is necessary to know how to distinguish the pyroxenes from the amphiboles. The easiest method is to note the angle between the cleavage faces. In pyroxenes the good, but not perfect, cleavage faces meet at nearly right angles (87° and 93°), whereas in amphiboles the highly perfect cleavage faces meet at angles of 55° and 125° . Thus, the cleavage prism is nearly square in pyroxene and distinctly oblique in amphibole.

Augite is the most important of the pyroxenes. It does not have a constant composition but is essentially a metasilicate of calcium, magnesium, and ferrous iron, together with silicates of ferric iron and aluminum; manganese and alkalis are often present. It has a hardness of 5.0–6.0 and a specific gravity of 2.93–3.49. It can frequently be identified in rocks because of its dull cleavage face, black color, and cleavage angle approximating 90° . Augite occurs in many types of igneous rocks and is an

essential constituent of diabase and basalt. Although it alters to many different minerals, it is more stable than hornblende and may occur as "black gravel" in some soils.

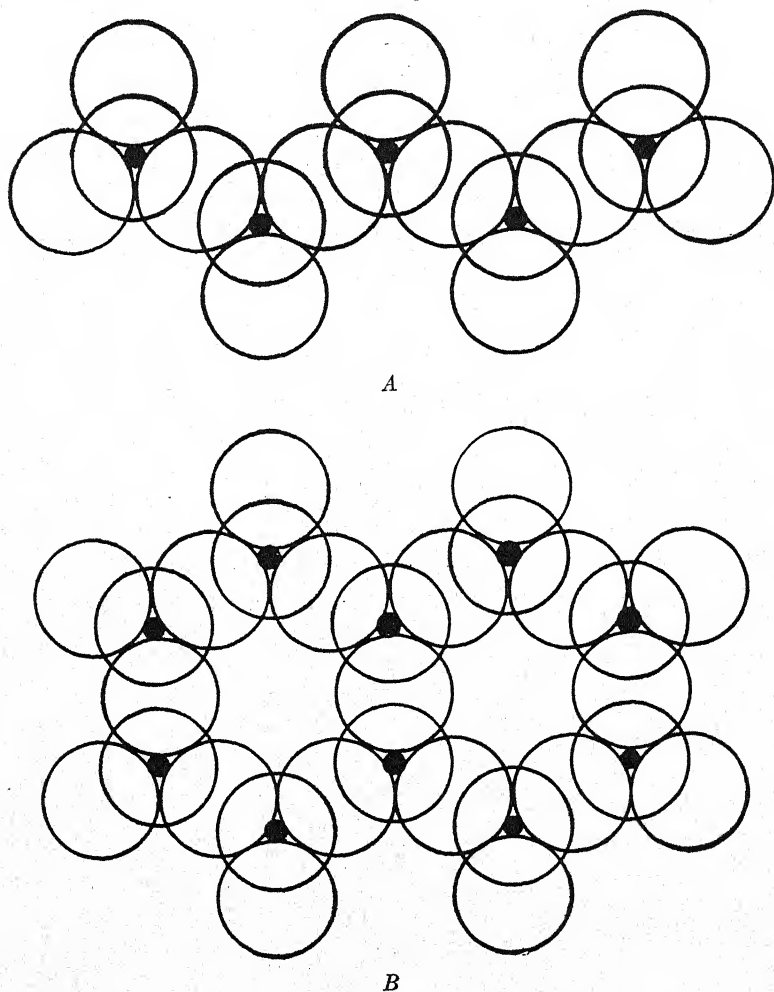


FIG. 2. A, a single chain of oxygen tetrahedra occurring in the pyroxenes; B, double chains of tetrahedra occurring in the amphiboles.

Other less important pyroxenes are *enstatite*, MgSiO_3 , and *hypersthene*, $(\text{Mg,Fe})\text{SiO}_3$. Both occur in basic eruptive rocks and may alter to talc and serpentine as a result of hydration.

Hornblende, the most important of the amphiboles, has a variable composition; compared with augite, it contains a smaller proportion of calcium but more magnesium and iron. It is usually quite black in color and can be distinguished from augite by its highly perfect cleavage faces, which meet at angles of 55° and 125° . It has a hardness of 5.0–6.0 and a specific gravity of 3.0–3.47. Hornblende is widely distributed, occurring in granite, syenite, diorite, diabase, gabbro, and norite and in the metamorphic gneisses, hornblende schists, and amphibolites. The mineral decomposes readily and forms such alteration products as pyroxene, chlorite, epidote, biotite, siderite, calcite, quartz, and clay. Its rate of weathering in soils is often used as an indication of the intensity or duration of soil-forming processes in cool, humid regions.

Amphiboles which are of lesser importance but which merit mention are *tremolite*, $\text{CaMg}_3(\text{SiO}_3)_4$, and *actinolite*, $\text{Ca}(\text{Mg,Fe})_3(\text{SiO}_3)_4$. Tremolite is most commonly encountered in dolomitic limestones, actinolite is often abundant in crystalline schists.

The pyroxenes and amphiboles are an original source of calcium, magnesium, and iron in soils.

MICAS

The micas are aluminosilicates containing hydroxyl. Their ionic structure is closely related to that of the important group of clay minerals, montmorillonite. The basic structure consists of a sheet of silicon-oxygen tetrahedra (Figure 1C) linked together at each corner. This sheet is linked by means of common oxygen ions to a sheet of aluminum-oxygen octahedra directly above, which in turn is linked to another sheet of silicon-oxygen tetrahedra. The result is a sheet-like structure with three layers. This structure causes the micas to have perfect basal cleavage and accounts for the flake-like nature of the minerals. There are two micas which concern the soil scientist: muscovite, the more or less colorless variety, and biotite, the black mica.

Muscovite, also called potash mica, has the formula $\text{KAl}_2(\text{AlSi}_3\text{O}_{10})(\text{OH})_2$. The expression $(\text{AlSi}_3\text{O}_{10})$ represents the composition of the silicon-oxygen tetrahedral sheets on the top and bottom of the three-layered structure. The silicon-

oxygen tetrahedral sheets would normally have the formula Si_4O_{10} , but one silicon has been replaced by aluminum. The Al_2 and $(\text{OH})_2$ are located between the two silicon-oxygen sheets, with the aluminum in six-coordination (octahedral arrangement), and the OH groups occupying the same spaces as would an equal number of oxygen ions.

Muscovite is fairly soft, ranging from 2.0 to 2.5 in the scale of hardness. Primary muscovite is essentially a mineral of granites, quartz porphyries, and mica schists. It is fairly abundant in gneisses but is less commonly associated with the more basic rocks. Muscovite is readily recognized in rocks because of its perfect basal cleavage, flaky form, and transparent to translucent properties. It is a reserve source of potassium supply in soils. Denison *et al.* (1929) found muscovite in Piedmont soils to vary greatly in potash content, ranging from less than 1 per cent to as much as 9 per cent. This variability is undoubtedly explained by the fact that, upon weathering, the potassium, which is located between the crystal plates, is hydrolyzed or replaced by other metallic cations to varying degrees.

Biotite, or black mica, has the same general structure as muscovite except that the aluminum located between the two layers of silicon-oxygen tetrahedra is replaced by magnesium and iron. Therefore the idealized formula for biotite is $\text{K}(\text{Mg,Fe})_3(\text{AlSi}_3\text{O}_{10})(\text{OH})_2$. It is interesting to note that the basic structural part of the formula does not change. It may appear confusing that $(\text{Mg,Fe})_3$ in biotite has replaced Al_2 in muscovite. The reason, according to Bragg (1937: 211), is that in muscovite only two-thirds of the available places of six-coordination between O and OH groups have been used. Therefore it is possible in biotite to use all the spaces and to substitute Al_2 with $(\text{Mg,Fe})_3$. Because of differences in valence a balanced structure results in both minerals.

Biotite is readily recognized in rocks because of its black color, bright cleavage faces, and flaky character. It is a common constituent of such massive igneous rocks as granite, syenite, diorite, trachyte, and andesite. This mineral is more generally distributed than muscovite and is considered to be more subject to chemical alteration. The end products of alteration may be chlorite, or more rarely serpentine, and accompanying kaolinite. The formula for biotite indicates that it contains almost as much

potassium as muscovite; in addition it contains magnesium and iron, both of which are essential plant nutrients. Perhaps it was these differences which caused Ramann (1905: 78) to state that soils derived from rocks containing much biotite are rich in iron and clay material and because of their physical properties and higher productivity can be distinguished from soils derived primarily from muscovite.

CHLORITES

The chlorites are silicates of aluminum with magnesium and iron. They are similar to the micas in having very good basal cleavage, but differ in being inelastic. The chlorites are more basic than the micas, are highly hydrated, and are free, or nearly free, from alkalis. The ionic structure of chlorite is very interesting, since it consists of alternating mica-like and brucite-like layers. The mica-like layers have a composition ranging between $Mg_3(AlSi_3O_{10})(OH)_2$ and $Mg_2Al(Al_2Si_2O_{10})(OH)_2$, whereas the brucite-like layers have the formula $Mg_2Al(OH)_6$. The aluminum and magnesium in the brucite-like layers are, of course, in the octahedral position with six shared OH ions packed around each cation.

Chlorites are always of secondary origin and are abundant in metamorphic schists, where they commonly occur as scaly aggregates. They can usually be identified in rocks by their dull greenish color and their thin flaky form; also they are rather soft, having a hardness of 2.0-2.5. The chlorites disintegrate rather readily, but chemical decomposition is slow. They are not an important factor in soil fertility, for magnesium is the only plant nutrient contained, and this element is usually present in more readily available form in other soil minerals. According to Hilgard (1906: 36), chlorite schists frequently give rise to inferior soils.

OLIVINE GROUP

The olivines are orthosilicates of iron and magnesium. The minerals forsterite, Mg_2SiO_4 , and fayalite, Fe_2SiO_4 , are rare, but they commonly form an isomorphous mixture known as *olivine*, $(Mg,Fe)SiO_4$. Heide (1929) stated that the ratio of Mg:Fe

varies from about 16:1 to 2:1. The hardness is 6.5–7.0, and the specific gravity 3.2–3.6. The SiO_4 in olivine is present as independent silicon-oxygen tetrahedra. The magnesium and iron occur between these tetrahedra surrounded by six oxygen ions. These oxygen ions are part of the tetrahedra, and room for the other cations occurs because every other tetrahedron is inverted. Olivine is an essential constituent of many eruptive rocks, such as peridotite, norite, basalt, diabase, and gabbro. Dunite is a rock consisting of olivine alone. As a rule olivine occurs only in rocks low in silica. It alters readily, yielding such minerals as serpentine, magnetite, magnesite, calcite, and quartz; limonite, epidote, and clay are other alteration products. In western Norway olivine, particularly varieties rich in magnesium (Mg:Fe about 9:1), appears to give rise to good soils for Scotch pine (*Pinus sylvestris*)³ (Nilsen, 1936). According to Burt (1927), favorable soil conditions resulting from the presence of olivine in the parent material are to be explained by the fact that the mineral weathers readily, thus facilitating disintegration of the rock into soil.

Olivine can usually be identified in rocks by its green color, appreciable hardness, indistinct cleavage, and conchoidal fracture. It may be distinguished from greenish pyroxene by its greater hardness and lack of well-defined cleavage.

SERPENTINE

Serpentine has the chemical composition $\text{H}_4\text{Mg}_3\text{Si}_2\text{O}_9$ and a hardness of 2.5–3.0. It occurs only as a secondary mineral resulting from the alteration of olivine, hornblende, enstatite, actinolite, and other magnesian minerals. Gabbro, peridotite, and amphibolite may undergo alteration to impure serpentine. Serpentine can usually be identified in the field by its mottled green color, greasy or waxy feel, and softness.

Serpentine is of little value as a soil-forming mineral. In fact, soils which are derived primarily from serpentine rock are

³ Where the authority is not indicated, the scientific names follow the nomenclature of *Standardized Plant Names*, second edition, 1942. As a rule scientific names are given the first time a species is mentioned but are not repeated thereafter. A complete list of common names and scientific names of trees mentioned may be found in the Appendix, pp. 486–487.

commonly shallow and infertile and have inferior physical properties. Such areas are often referred to as serpentine barrens. Hilgard (1906: 36) stated that soils derived from both serpentine and talcose rocks are almost always poor in nutrients and may be practically sterile. Gordon and Lipman (1926) investigated California soils derived from serpentine and other magnesian rocks and concluded that their infertility was caused by an alkaline reaction and a deficiency in certain ions, chiefly nitrate and phosphate, rather than by too high a content of soluble magnesium. It is unfortunate that these workers did not carry out their studies with actual soils rather than entirely with extracts from soils derived from magnesian rocks. Wherry (1932) investigated the magnesium content of a number of native plants growing on serpentine barrens in eastern Pennsylvania and found that the magnesium content of every plant was more than 100 per cent higher than that of the same species growing on normal Piedmont or coastal plain soils. He did not, however, demonstrate any toxic effect of this high content of magnesium. Novák (1937) noted that the vegetation of soils derived from serpentine frequently is quite different from that of neighboring soils derived from other parent materials.

TALC

Talc has the chemical composition $\text{Mg}_3(\text{Si}_4\text{O}_{10})(\text{OH})_2$. It is interesting to note the similarity between talc and mica from the standpoint of ionic structure. It may be recalled that the linked tetrahedral sheets in mica have the formula $\text{AlSi}_3\text{O}_{10}$, which means that one tetrahedral silicon has been replaced by aluminum. In talc, however, these tetrahedral sheets have the composition Si_4O_{10} , and the magnesium and OH groups occur between the two tetrahedral layers. The whole unit is neutral; hence no alkali or other cations are required between the plates, as they are in mica, to balance the negative charge. This fact, as will be pointed out in Chapter 7, is extremely important in considering the properties of clay minerals.

Talc is common as a pseudomorphous mineral, derived from other magnesian species, such as tremolite or enstatite. It has a hardness of only 1.0 and a specific gravity of 2.7-2.8. Outstanding characteristics which make it easy to recognize include

a greasy feel, extreme softness, pearly luster, and the fact that it is sectile. Talc frequently may be associated with marble or dolomite, and it forms talcose schists. Because of its softness and easy cleavage it is readily abraded and disintegrated; chemical alteration is slow. It forms a minor and rather inert constituent in soils.

EPIDOTE

Epidote has the composition $\text{Ca}_2(\text{Al,Fe})_3(\text{OH})(\text{SiO}_4)_3$. Thus it is a basic orthosilicate of calcium and aluminum, with varying replacement of aluminum by ferric iron. Its hardness is 6.0–7.0, and it has a specific gravity of 3.25–3.50. The ionic structure has not been worked out. Epidote is a common mineral in granites, schists, gneisses, and amphibolite. It also abounds as a secondary mineral derived from feldspars, pyroxenes, amphiboles, garnet, and biotite and is frequently associated with chlorite. The general color is usually green, varying from yellowish green to almost black, according to the amount of iron present. Distinguishing characteristics include the peculiar greenish color, marked hardness, and perfect cleavage in one plane only. Epidote is very resistant to chemical weathering. It breaks up slowly in soils, yielding calcite, quartz, kaolinite, and iron oxide. Generally it is more abundant in mature than in immature soils.

OXIDES, HYDROUS OXIDES, AND HYDROXIDES

The general ionic structure of the minerals considered in this section conforms to the same principles that were given for the silicate minerals. In quartz, for example, the silicon is in tetrahedral arrangement with each oxygen ion shared by two silicon ions; the tetrahedral groups are linked by common corners. The actual arrangement of the tetrahedra varies with each form of quartz, such as α - and β -quartz, and the high-temperature polymers tridymite and cristobalite.

The common sesquioxides, Al_2O_3 and Fe_2O_3 , of course, are in closest hexagonal arrangement with two successive layers of oxygen ions, between which the cations occur surrounded by six oxygen ions. If all positions were filled, there would be as many cations as oxygen ions. Actually only two-thirds of the available positions are occupied.

The hydroxides are similarly formed, since the hydrogen ion is so small that it can be considered as occupying no space. In hydrargillite, $\text{Al}(\text{OH})_3$, for example, only two-thirds of the positions are occupied between the two oxygen sheets, whereas in brucite, $\text{Mg}(\text{OH})_2$, all the available positions are filled with magnesium ions.

Quartz (SiO_2) has a hardness of 7.0 and a specific gravity of 2.65. Next to the feldspars, quartz is the most abundant mineral in the earth's crust. It forms about 12 per cent of the entire lithosphere (Clarke, 1924). It is an important constituent of granites and pegmatites and is even more important in secondary rocks, such as schist, gneiss, quartzite, and sandstone. Because of its hardness, low degree of solubility, lack of cleavage, and resistance to decomposition it is a common and usually abundant constituent of the coarser fractions of soils. Coffey (1913: 97), in reporting on a mineralogical examination of the surface soil (sand and silt fraction) from twenty-five different soil series, stated that quartz is by far the most common mineral. In general, it is most abundant in soils derived from acidic rocks and in those where the material has undergone the most weathering. Quartz may dissolve but does not decompose chemically; silicates, however, are commonly decomposed by ground waters. Although quartz is perceptibly soluble in alkali-carbonate solutions, it is nearly insoluble in the acids of ground waters. Quartz is readily recognized in rocks by its glassy appearance, great hardness, and conchoidal fracture. Its contribution to soils consists largely of influencing soil texture, soils high in quartz being coarse-textured.

Limonite is a hydrated ferric oxide having the chemical composition $2\text{Fe}_2\text{O}_3 \cdot 3\text{H}_2\text{O}$. Commonly the mineral carries considerable amounts of phosphorus and manganese as impurities. It is one of the very common soil constituents and one of the most important coloring agents of brown and yellow soils. It may serve as a cementing material in hardpans, but not as commonly as does *turgite*, $2\text{Fe}_2\text{O}_3 \cdot \text{H}_2\text{O}$. On dehydration limonite forms hematite; on reduction and carbonation it yields soluble iron. Limonite is a common alteration product of pyrite, marcasite, hornblende, and pyrrhotite. In northern latitudes limonite usually occurs as finely divided matter coating rock and soil particles, but in southern and tropical soils it, along with hematite and the

oxides and hydroxides of aluminum, may constitute a large portion of the soil material.

Hematite (Fe_2O_3) occurs in the siliceous and feldspathic granites, syenites, trachytes, rhyolites, andesites, and phonolites, as well as in crystalline schists. Hematite alters to limonite, magnetite, pyrite, marcasite, and siderite. It occurs as coatings on sand grains, as incrustations, and as pulverulent matter disseminated throughout the soil body. It may occur as the cementing material in hardpans; with turgite, it is one of the most important causes of the red color in soils. Perhaps the most convenient means of distinguishing hematite from limonite is the streak; that of hematite is red, and that of limonite, yellow.

Magnetite ($\text{FeO} \cdot \text{Fe}_2\text{O}_3$), is a heavy primary mineral occurring in rocks of all classes; its specific gravity is 5.17. Occasionally magnetite may be present in soils as black sand, but as a general rule it is of little direct importance because of its scarcity as a soil mineral. It usually can be identified in rocks by its metallic luster, hardness (5.5-6.5), strong magnetism, and octahedral parting.

Gibbsite (hydrargillite), $\text{Al}(\text{OH})_3$, and *diaspore*, $\text{AlO}(\text{OH})$ or $\text{Al}_2\text{O}_3 \cdot \text{H}_2\text{O}$, are the two most common aluminum compounds in soils. They occur as coatings on the larger mineral particles in the accumulative horizons of podzolic soils, and they represent a large portion of the volume in tropical soils. The pure minerals, which are white, are seldom seen in nature. In tropical soils their natural color is masked by the red and yellow of iron compounds.

PHOSPHATES

Phosphorus is one of the mineral nutrients that is commonly deficient in soils. The availability of phosphorus to plants is, to a considerable extent, dependent upon the form in which it is present. The most readily available source is that in the calcium and magnesium phosphates. Compounds of iron and aluminum and phosphorus occur, particularly in acid soils; these compounds are relatively stable at low pH values.

The most important primary source of phosphorus in the soil is the mineral *apatite*. There are two common forms, *fluorapatite*, $\text{Ca}_5(\text{PO}_4)_3\text{F}$, and *chlorapatite*, $\text{Ca}_5(\text{PO}_4)_3\text{Cl}$. The hard-

ness is 5.0, and the specific gravity 3.17–3.23. Apatite occurs in all classes of rocks, but it is more common in femic rocks than in highly siliceous varieties. Except for the large deposits of phosphate rock that are mined for fertilizer, apatite seldom occurs in large amounts, but it is widely disseminated as very small crystals in rocks. Apatite, compared to many other rock minerals, decomposes rather rapidly under the influence of water containing carbonic acid, both calcium and phosphorus passing into solution. Field identification of apatite occurring as minute crystals in rocks is scarcely possible.

CARBONATES

The most important carbonate mineral in soils is *calcite*, CaCO_3 . Its hardness is 3.0, and its specific gravity 2.72. Calcite alone, as a rock, is represented by marble, limestone, and chalk. It occurs most abundantly in sedimentary rocks in combination with other components. It is readily altered by the action of carbonated waters to the bicarbonate, $\text{CaH}_2(\text{CO}_3)_2$, and may completely dissolve out of limestone, leaving only the impurities behind. The mineral is widespread in occurrence and is the principal direct source of calcium in soils. Of the various soil minerals calcite probably is the most important, for it influences soil acidity and accordingly the distribution and growth of plants. It can be readily detected in all rocks and soils by its effervescence in cold dilute acids.

Dolomite has the chemical composition $\text{CaMg}(\text{CO}_3)_2$ but is usually impure. The calcium of limestone may be partially replaced by magnesium to form dolomite. This replacement leads to a shrinkage in volume of the original rock mass with the development of cavities and pores. Dolomite is less readily decomposed than calcite; but, like limestone, it usually decomposes, leaving the impurities behind to form the soil. Dolomite is a very important source of magnesium in soils. Unlike calcite, dolomite does not effervesce in cold dilute acid; in hot acid it is readily soluble with effervescence.

Siderite has the chemical composition FeCO_3 . It is a common alteration product of other iron-containing minerals and may itself alter to hematite, limonite, or turgite. It may be of importance in the formation of hardpan in certain soils.

SULFATES

The only sulfate to be considered here is *gypsum*, $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$. It occurs abundantly in some of the drier regions; as an important rock it is generally a saline residue. Gypsum is deposited in the lower horizons of some soils in the semiarid to arid regions. Its primary use in the eastern humid region is in the manufacture of fertilizer and as a soil amendment. The distinguishing characteristics of gypsum are its perfect cleavage in one direction and its extreme softness (2.0).

SULFIDES

No discussion of common minerals would be complete without *pyrite*, FeS_2 . It has a hardness of 6.0–6.5 and a specific gravity of 4.95–5.10. Pyrite is a common accessory mineral in igneous rocks. It may alter to limonite, hematite, goethite, or iron sulfate. This mineral, which is commonly called fool's gold, can be distinguished by its pale, brass-yellow color and conchoidal to uneven fracture. Its ready alteration to injurious ferrous sulfate and sulfuric acid causes it to be an undesirable mineral in some soils.

THE MINERAL COMPOSITION OF SOILS

The mineral composition of soils is influenced by a number of factors, chief among which are the nature of the parent material, the pedogenic processes which have been operative, and the age of the soil.

It appears that soils usually contain a greater variety of minerals than do rocks (Coffey, 1913: 96). Mineralogical analyses of American soils made by Cameron and Bell (1905), Failyer, Smith, and Wade (1908), McCaughey and Fry (1913), and Robinson (1914) indicate that most of the common rock-forming minerals can be identified in every soil. The occurrence of a variety of minerals in most soils may be explained by the fact that the parent materials are usually derived from not one but several kinds of rock.

It must not be assumed, however, that the mineralogical composition of soils is usually the same. Very great differences occur in the *proportion* of different minerals in different soils. Parent materials derived in large part from granite give rise to

soils containing a larger proportion of quartz than do parent materials derived from diorite. In general, there appears to be a fair relation between the mineralogical composition of soil parent materials and the composition of the underlying rocks. This relationship which is to be expected in residual soil materials, also appears to exist in many transported soil materials. Bowman (1911: 660) stated that analyses made by Leverett and Alden showed about 85 per cent of the glacial till in the Great Lakes region to be derived from the underlying sedimentary rocks. Bowman believed that in the Connecticut Valley the proportion of locally derived material was even higher. The relationship between glacial drift and the adjacent underlying rocks is usually closer where the drift is thin than where it is thick (Emerson, 1920). Investigators in other countries also have found that the mineralogical composition of soils derived from transported materials is strongly influenced by the composition of the underlying rock (Hendrick and Newlands, 1923, 1925; Hart, 1929a, 1929b; Elder and McCall, 1936).

Cameron and Bell (1905) in their investigation of the mineralogical composition of many American soils found that as a rule the finer the particles were, the lower was the proportion of quartz and the higher the proportion of other minerals. This same situation was reported by Thomas (1923) as occurring in Hagerstown silty clay loam soil.

The influence of pedogenic processes on the mineral composition of soils is recognized, but specific information is scanty. It is to be expected that soils developed in climatic zones where chemical weathering is intense, as in regions of podzol soils, will contain a lower proportion of readily altered minerals and a higher proportion of minerals resistant to alteration than will soils developed under conditions of less intense weathering. In the intensely weathered *A* horizon of podzol soils there is commonly a much higher percentage of minerals resistant to alteration, such as quartz and orthoclase, than is found in the less weathered *B* and *C* horizons (Tamm, 1934, 1937; Semb, 1937). In brown forest-soil profiles, where leaching has been less intense, the mineral composition of the different horizons tends to be more nearly the same (Semb, 1937). Cady (1940) and Richard and Chandler (1944) have presented evidence to indicate a marked reduction in the amount of certain heavy minerals, particularly hornblende,

in the A horizon of podzol soils. The difference is not so great in the brown podzolic soils. Podzol and brown podzolic forest soils which have been disturbed by the uprooting of trees show a significantly higher proportion of minerals of high specific gravity than do adjacent undisturbed soils (Lutz, 1940). Minerals having a specific gravity of < 2.680 are commonly referred to as light minerals; examples are quartz, orthoclase, albite, and oligoclase. Minerals having a specific gravity > 2.680 are commonly referred to as heavy or basic minerals; examples are labradorite, anorthite, augite, hornblende, muscovite, biotite, and apatite. The minerals in the first group generally weather slowly; those in the second group generally weather more readily.

The age and maturity of a soil have a bearing on its mineral composition. With increasing age and weathering of soil material the proportion of readily altered minerals decreases, and the proportion of minerals resistant to alteration increases. As a soil body becomes more mature, its mineral composition departs more and more from that of the parent material; primary minerals are broken down and, in part at least, replaced by secondary minerals representing alteration products.

Knowledge of the mineral composition of soils is valuable from several points of view. Marshall and Haseman (1943) and Mickelson (1943) have demonstrated that information on heavy minerals resistant to weathering may be used in the quantitative measurement of soil formation. Mineral composition is also very useful in determining the source of the parent material. Edelman (1936) recognized two broad classes of soil profiles based on the source of the minerals contained; (1) autochthonous, containing only minerals which were originally present at the spot or which were derived from these as secondary products, and (2) allochthonous, containing only minerals that were transported to the spot, not material from the underlying rock. Between these two extreme situations are all gradations. Information concerning the mineral content may be utilized in the classification of soils (Hart, 1929a, 1929b; Edelman and van Beers, 1939; Jeffries and White, 1940). The results of investigators in various parts of the world indicate that the mineralogical composition of soils also has considerable ecological significance. In 1913 Coffey suggested that soils having a variety of minerals may, as a rule, be more productive than those having a less varied com-

position. He remarked, "While there are some exceptions it may be stated that, in general, the soils having the lowest percentage of quartz are the most productive. Much depends, of course, upon the nature of the other minerals present, but it is nearly always true when comparing soils derived from the same general group of rocks."

Work by Tamm (1934, 1937) and Edelman and van Beers (1939) has substantiated the view of Coffey. Tamm found that forest growth in Sweden is almost always poor on soils containing a low proportion of heavy minerals. Somewhat earlier, Albert (1928: 117) expressed the view that soils rich in minerals supplying nutrient elements are better able to carry pure stands continuously than are soils which are poorer in these minerals.

In recent years considerable attention has been devoted to methods appropriate for the investigation of minerals in soils. It is not possible here to do more than mention some of the investigators who have concerned themselves with this field of inquiry. Among those who have worked with the minerals in the noncolloidal fractions are Krönig (1914), Short (1931), Steinriede (1931), Larsen and Berman (1934), Fry (1933), Tamm (1934, 1937), and Pearson and Truog (1939). The method of Tamm, by which the minerals in the particle-size class 0.6–0.2 mm. are separated into two groups (specific gravity < 2.680 and specific gravity > 2.680), is well adapted to investigations of forest soils. The identification and investigation of clay minerals have been carried on by Marshall (1930), Hendricks and Fry (1930), Kelley, Dore, and Brown (1931), Kelley, Jenny, and Brown (1936), Agafonoff (1936), Edelman, van Baren, and Favejee (1939), Hardon and Favejee (1939), Correns (1940), and others to be mentioned in Chapter 7.

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Chapter 3

SOIL-FORMING ROCKS

Rocks may be defined as naturally occurring solid materials forming a considerable part of the earth's crust. As a general rule, they consist of two or more minerals, but notable exceptions occur. For example, pure limestone or marble consists of calcite only; and obsidian, resulting from rapid cooling of molten material, is undifferentiated glass lacking definite minerals.

In view of the importance of rocks from the standpoint of geomorphology and soil development it is essential that the student gain a working knowledge of their mineral composition and classification. Although this background is best obtained in formal courses in petrology, it has seemed desirable to present here, for the benefit of those with limited training in geology, a brief treatment of the rudiments of the subject. For a more complete presentation a standard textbook on petrology should be consulted.

All rocks are conveniently assigned to one of three broad groups: igneous, sedimentary, and metamorphic. *Igneous* rocks are formed as a result of solidification of molten material (magma) without any appreciable subsequent change. Granite and basalt are examples of common igneous rocks. *Sedimentary* rocks result from the deposition of sediments, either inorganic or organic in origin, in water; in addition, certain sedimentary rocks are of aeolian origin. Conglomerate, sandstone, shale, and limestone may be mentioned as examples of common sedimentary rocks. *Metamorphic* rocks result from secondary action, such as heat, pressure, or chemical action of liquids and gases, on igneous or sedimentary rocks. Thus, granite subjected to heat and pressure may be altered to gneiss, and a sandstone infiltrated with silica becomes a quartzite.

According to Clarke and Washington (1924), the composition of the earth's crust, with reference to the broader rock groups, may be estimated as follows: igneous rocks, 95 per cent; shales,

4.0 per cent; sandstones, 0.75 per cent; and limestones, 0.25 per cent.

IGNEOUS ROCKS.

To be most useful to the student of forest soils a classification of igneous rocks must be practical in the sense that it is based on megascopic features which can be observed in the field. Fortunately there are a number of classifications which fulfill this requirement, being based primarily on evident characteristics, such as texture and mineralogical composition. The classification presented by Grout (1940) is well suited to the needs of students of forest soils and is here presented in adapted form as Table 4.

Some explanation of the use of Table 4 is necessary. It will be noted that the texture designations in the second vertical column range from granitoid (coarse crystals, all about the same size) at the top, to felsitic (very fine or microscopic crystals) and glassy (like glass, no definite minerals) at the bottom. These texture differences are related to the mode of occurrence, which is indicated in the first vertical column. Thus rocks having large crystals were formed in deep-seated positions where cooling of the magma was slow, whereas rocks with very fine crystals (or without crystals, as in obsidian) were formed in surface positions where cooling was very rapid. Occupying an intermediate position between the rocks of granitoid and of felsitic texture are those exhibiting porphyritic texture. In general this texture is developed in small intrusive bodies, where cooling is more rapid than in deep-seated positions and less rapid than in surface positions. Characteristically the porphyritic texture features large crystals, called *phenocrysts*, embedded in a finer *groundmass*. If the groundmass is so fine that individual minerals cannot be distinguished megascopically, the presence of well-developed phenocrysts commonly provides a basis for identification of the rock. Examples of various types of texture are presented in Figures 3, 4, 5, and 6.

It is important to recognize that a magma of given chemical composition may give rise to several different kinds of rocks, depending on the physical conditions under which solidification occurs. Thus granite, granite-porphyry, rhyolite-porphyry, and rhyolite have essentially the same chemical composition, although

on the basis of texture they are different kinds of rock. This relationship of rock types within each of the vertical columns of Table 4 should be borne in mind.

The primary divisions across Table 4 are based on the most abundant feldspar, with subdivisions according to the presence or lack of such minerals as quartz, olivine, biotite, hornblende, and augite. It should be mentioned that a minus sign preceding certain minerals is not to be interpreted literally; the intention is

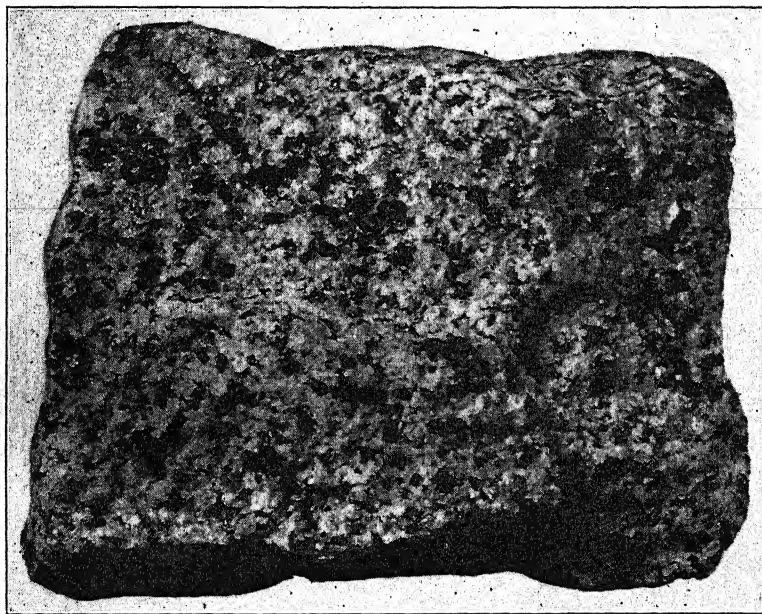


FIG. 3. Granite, illustrating granitoid texture. Reprinted by permission from *Outlines of physical geology*, by C. R. Longwell, Adolph Knopf, and R. F. Flint, second ed., 1941. John Wiley and Sons, Inc. Photograph by United States Geological Survey.

to indicate that the mineral is not present in notable amounts. Thus, a granitoid rock having plagioclase as the predominant feldspar and containing biotite or hornblende may be a diorite even though it contains a small amount of quartz, perhaps 2-3 per cent.

It is obvious that successful application of this classification is contingent on an understanding of rock textures and an ability

to distinguish with a hand lens the more common minerals. It is suggested that the student practice using the classification in Table 4 or a similar one, starting with a small but representative collection of authentic specimens of rocks and minerals. Soon it

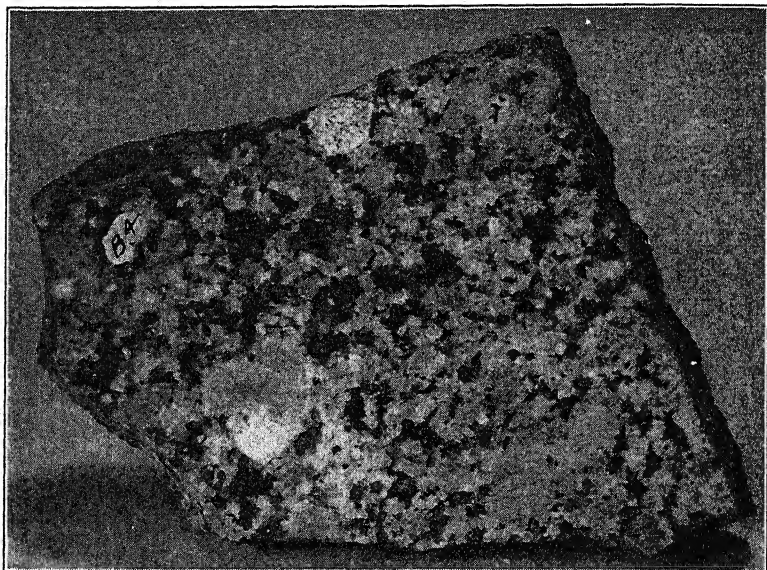


FIG. 4. Porphyritic granite, showing scattered large phenocrysts in a granitoid groundmass.

will be found that with the aid of a good hand lens the classification can be used directly in the field.

RHYOLITE-GRANITE CLAN

Members of this rock clan may be identified by the large proportion of light-colored minerals, such as orthoclase, plagioclase, and quartz.

A *granite* is a rock having a granitoid texture and containing the *essential* minerals just mentioned. In an average granite, orthoclase occupies about 50 per cent, plagioclase about 20 per cent, and quartz over 20 per cent of the rock. The dark-colored minerals biotite, hornblende, and augite (of which biotite is the most common) are present in lesser amounts and may be referred to as *accessory* constituents. The presence or absence of acces-

sory minerals does not fundamentally affect the nature of the rock, but the predominant one is referred to as the characterizing accessory and is used in the name. For example, biotite is usually the most common accessory mineral in granite; and, when it is present in an amount sufficient to give character, the rock is called a *biotite granite*. Similarly, when hornblende predom-

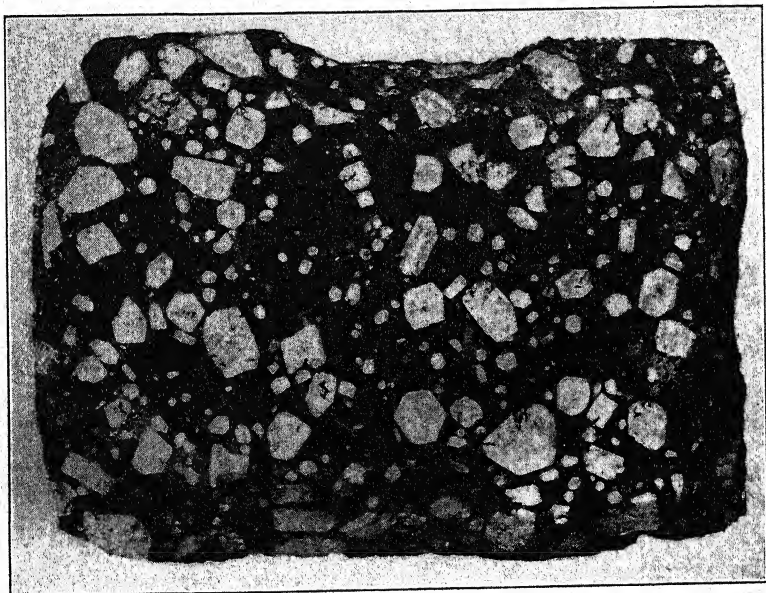


FIG. 5. Porphyry, illustrating porphyritic texture; the large, light-colored phenocrysts are inclosed in a fine-grained groundmass. Reprinted by permission from *Outlines of physical geology*, by C. R. Longwell, Adolph Knopf, and Richard F. Flint, second ed., 1941. John Wiley and Sons, Inc.

inates, the rock is called a *hornblende granite*. Seldom do the dark-colored minerals represent more than 25 per cent of the volume of rocks in this clan; as a rule, they are present in considerably smaller quantities.

The predominance of large phenocrysts of feldspar and quartz in a groundmass of felsitic texture indicates a *granite-porphyry*. Prominent phenocrysts (constituting perhaps one-tenth to one-half the rock) in a groundmass having a more finely felsitic texture characterize the rock as a *rhyolite-porphyry*. The chief difference between a *granite-porphyry* and a *rhyolite-porphyry* is

that the granite-porphyry has larger and more abundant phenocrysts in a groundmass which, although felsitic, is slightly coarser. If phenocrysts are lacking or very scarce and the texture is felsitic, the rock is a *rhyolite*, providing its chemical composition is similar to that of the other members of this class. Actually it is not possible by megascopic means to distinguish rhyolites from

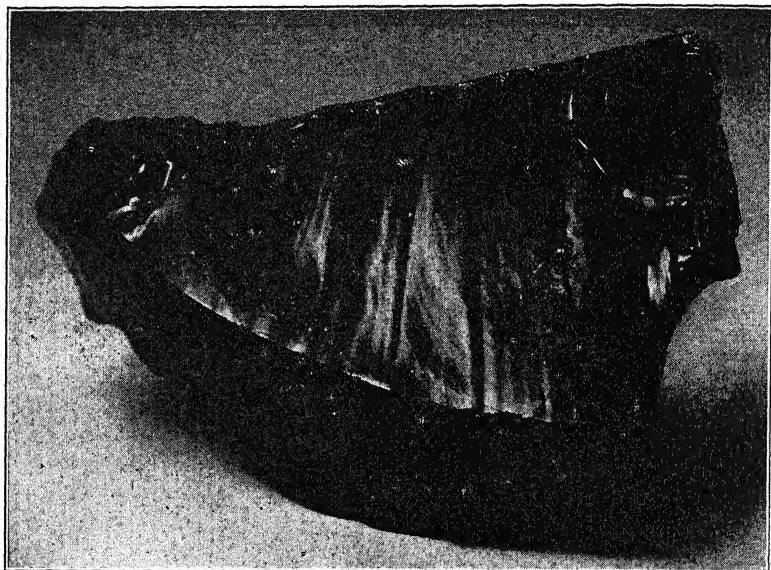


FIG. 6. Obsidian, illustrating glassy texture.

trachytes and dacites when phenocrysts are lacking. In view of this fact, light-colored felsitic rocks are frequently designated as *felsites*, a term which has the advantage of being noncommittal.

Rhyolites do not give rise to very productive soils. This point was stressed by Hilgard (1906: 53). The generally poor quality of soils derived from rhyolites seems to be associated with the resistance of the rocks to weathering; soils from this type of parent material contain a large amount of quartz and relatively undecomposed glass. In the Sierra Nevada Mountains of California there are areas of considerable extent with soils derived from volcanic glass. Usually they support only poor stands of trees. Show (1930) has pointed out that repeated efforts to establish

sugar pine (*Pinus lambertiana*) on the so-called lava-ash soils have resulted in nearly universal failure. These soils are very loose and porous, and sugar pine does not occur on them naturally.

Granites weather to more productive soils than do rhyolites. In general, the coarser-grained a granite is, and the richer it is in feldspars, the more readily and deeply it weathers. Bowman (1911: 607) noted that in the southern Appalachian Mountains the feldspathic Cranberry granite has weathered much more rapidly than have rocks with lower feldspar content.

Soils of humid regions developed from parent materials arising from rocks of the rhyolite-granite clan are nearly always acid, because the calcium content of the rocks is low (usually less than 2 per cent). Furthermore, these soils tend to be sandy in texture because of their high content of quartz. These two features, low calcium content and sandy texture, explain the tendency of such soils toward development of mor humus layers¹ and podzolization, particularly in cool, humid regions.

The rhyolites occur principally in the western states, being common in the Black Hills of South Dakota, Yellowstone National Park, Colorado, Nevada, and California. Granites are widely distributed, being prominent along the Atlantic coast, in the Appalachian region, the northern Lake States, the Black Hills of South Dakota, the Rocky Mountains, and the Sierra Nevada Mountains.

TRACHYTE-SYENITE CLAN

Rocks in this group are closely related to the rhyolites and granites. The principal difference is that the *trachytes* and *syenites* contain no quartz, or at most not over 5 per cent. They are composed chiefly of feldspars, orthoclase being predominant. Dark-colored minerals, such as biotite, hornblende, augite, and magnetite, do not aggregate more than about 20 per cent. The mineral composition suggests, and chemical analyses confirm, a lower silica content and higher alumina and alkali content than in those of members of the rhyolite-granite clan. In mode of occurrence the trachytes are comparable to the rhyolites, and the syenites to the granites.

¹ A discussion of the nomenclature of forest humus layers will be found in Chapter 6.

The trachytes usually give rise to coarse-textured soils of low fertility (Ramann, 1905; Hilgard, 1906: 53). They tend to be rich in alkalis but poor in calcium and phosphorus. According to Ramann (1905), soils derived from syenite rocks in Germany are favorable for the development of hardwood forests. It would not be expected, however, that soils derived from syenites would show any appreciable superiority over those derived from granites. The only advantage of syenite derivation is a lower content of quartz in the coarse fraction of the soil.

Trachytes are well known in the Black Hills of South Dakota but are uncommon in most sections of America. Syenites are more abundant than the trachytes, occurring in the Adirondack Mountains, Maine, and New Hampshire, near Little Rock, Arkansas, and at various other localities.

DACITE-QUARTZ-DIORITE CLAN

Quartz-diorites are granitoid rocks containing plagioclase as the principal feldspar and notable amounts of quartz. In addition, dark-colored minerals such as biotite, hornblende, and magnetite may be present. Quartz-diorites are distinguished from granites by a high plagioclase content and from other granitoid rocks by the presence of quartz. *Dacites* are felsitic rocks which cannot be distinguished from rhyolites and trachytes in the field. Consequently, as has been mentioned, they are commonly referred to as felsites. The nomenclature of the other rocks of this clan is shown in Table 4. In this clan, as in those previously considered, the basis for distinguishing the various members is texture.

Information on the character of soils derived from quartz-diorite and dacite rocks is scanty. However, it may be expected that their general productivity will at least equal that of soils from rocks of the rhyolite-granite clan.

Quartz-diorites are known to occur near Peekskill, New York, and in Yellowstone National Park and the Sierra Nevada Mountains. Dacites frequently appear as subordinate members in regions where andesites are the principal rocks. They are of widespread occurrence in the western cordillera.

ANDESITE-DIORITE CLAN

Rocks in this clan are characterized by the practical absence of quartz and the occurrence of plagioclase feldspar which equals or exceeds in amount the dark-colored minerals (biotite, hornblende, and sometimes augite). The granitoid member of this clan is *diorite*, and the felsitic member is *andesite*. Porphyries occupy the same general relationships noted in the other clans.

In general, rocks of the andesite-diorite group may be expected to give rise to soils more fertile than those derived from rocks of the rhyolite-granite group. This assumption is based on their higher content of plagioclase feldspars and the resulting increase in calcium. Data on this point, however, are both scanty and apparently conflicting.

Andesites are of widespread occurrence in the western United States; they are important in Yellowstone National Park and form many mountain peaks in Colorado. Mount Hood, Mount Shasta, and Mount Rainier, together with other mountain masses in the western states, are chiefly andesite. Diorites also are widely distributed and are known to occur at many localities in the western United States.

BASALT-GABBRO CLAN

Rocks in this group are frequently difficult to classify because of the great variability which exists and the gradual transition from one rock type to another. Granitoid rocks consisting principally of plagioclase and augite may be termed *gabbro*. The relative proportion of plagioclase and ferromagnesian minerals varies widely; some rocks may consist almost entirely of plagioclase and others almost wholly of ferromagnesian minerals (augite, olivine, hornblende, biotite). On the feldspathic extreme are rock masses of large size consisting of nearly pure, coarsely crystalline labradorite; these rocks are termed *anorthosites*. On the ferromagnesian extreme are rocks which consist wholly of ferromagnesian minerals. Thus *pyroxenites* consist of pyroxenes (augite and hypersthene) alone, *hornblendites* of hornblende alone, and *peridotites* of pyroxenes plus olivine. These rocks are very high in magnesium and relatively low in silica. On alteration the pyroxenites and peridotites may give rise to *serpentine* rocks.

The term *dolerite* is frequently used to designate dark-colored, basic rocks which, because of their fine-grained nature, cannot be classified with certainty as either diorite or gabbro. The uncertainty results from the impossibility of determining megascopically whether the dark-colored mineral contained is hornblende or augite. *Diabase* is a term applied to rocks in the basalt-gabbro clan which have a characteristic pattern in their texture. In the coarse-grained members the pattern is formed by rods or laths of plagioclase surrounded by larger crystals of augite. It is impossible to distinguish by megascopic means the felsitic members of this clan. Consequently they are referred to as the *basalt group*. An analogous situation has already been mentioned in connection with the felsitic members of the clans of light-colored, acidic rocks. Gabbros occur at many points in the United States, notably in the White Mountains in New Hampshire, the Adirondack Mountains in New York, the Lake Superior region, and the West.

Temperate zone soils derived from gabbro are usually fertile and well supplied with calcium. Observations in Germany by Lang (1926: 233) and in Sweden by Tamm (1921) confirm this view. Anorthosites are very abundant in several Canadian areas and in the Adirondack Mountains, where Heimburger (1934) noted that the growth of red spruce (*Picea rubens*) is better in areas underlain by anorthosite than in sections mostly underlain by Adirondack gneiss. The ultrabasic rocks, hornblendite and peridotite, are commonly associated with the gabbros. Soils derived from these rocks, like those from serpentine, are usually low in productivity. They are generally shallow and poorly supplied with certain essential nutrients (Bowman, 1911: 147; Leiningen, 1912: 470; Merrill, 1921: 348; Tschermak, 1935; Domes, 1938: 473). Poor growth of forest trees on soils derived from ultrabasic rocks has been reported from the central Alps in Europe, the coast ranges and Sierra Nevada Mountains in California, southwestern Oregon, and Alaska. In the central Alps on soils derived from serpentine rocks, the timber line does not extend to as high altitudes as it does on soils derived from other parent materials (Leiningen, 1912: 470).

Diabases are well known in the eastern United States, where they occur as sills in the Triassic formations. The fertility of soils derived from these rocks is similar to that of soils from

other members of the basalt-gabbro clan. Ramann (1911: 588) stated that in the Mittelgebirge in Germany dikes of diabase may be recognized from a distance because of the better forest growth and the occurrence of exacting species, such as ash and maple. Soils derived from diabase are usually well supplied with calcium and phosphorus and produce good hardwood stands; reproduction is good, and mor humus layers seldom develop (Schmitt, 1926: 71; Niklas, 1930: 58-59).

Basaltic rocks are very common in North America. Soils derived from these rocks occur over large areas in the southwestern part of the United States; possibly half the saw timber in the Southwest is growing on soils having this origin. The greatest basaltic area in America is in the Snake River region of southern Idaho, extending into eastern Oregon and Washington. In general, the rate of weathering of basalt is slow, a fact which accounts for the commonly shallow, rocky nature of the resulting soils. If columnar jointing is well developed, the soils are apt to be very dry because of excessive drainage (Schuch, 1930: 259). On steep slopes basalt often forms talus; but, if moisture is plentiful, forest stands may develop. In general, soils derived from basaltic rocks are dark-colored and fertile; mor humus formation is rare (Ramann, 1905; Schmitt, 1926; Niklas, 1930: 62; Müller *et al.*, 1936: 734). Near New Haven, Connecticut, forest reproduction is readily established on soils of basaltic origin.

North and east of Flagstaff, Arizona, in the Coconino National Forest, is a considerable area covered to varying depth with black basaltic lapilli or "cinders" of comparatively recent origin. Establishment of ponderosa pine (*Pinus ponderosa*) reproduction on this material is difficult because of rapid drainage, low field capacity, and extremely high surface temperatures. When the material is more than 5-6 ft. deep, tree growth is lacking, but good ponderosa pine is obtained where a clay layer exists at a depth of 2-4 ft.

SEDIMENTARY ROCKS

SANDSTONE

Sandstones are formed when sands, a product of rock weathering, are reconsolidated because of pressure or the infiltration of cementing materials. In discussing the mineral composition of sandstones, Clarke (1924: 546-547) stated:

102009.

In short, all of the rock-forming minerals which can in any way survive the destruction of a rock may be found in its sands, and therefore in the sandstones. The feldspars and ferromagnesian minerals, however, are quite commonly altered or even removed, the more stable minerals, like quartz, being much more persistent. Quartz is the most abundant mineral in these rocks, while in rocks of the crystalline and eruptive classes it is subordinate to the feldspars.

Chemically, sandstones differ from sands chiefly because of the cementing material in sandstones. The cementing material may be silica, calcium carbonate, hydroxides of iron and aluminum, calcium sulfate, calcium phosphate, calcium fluoride, barium sulfate, or other compounds. The amount of cementing material in a sandstone may be very small or very large. Sandstones are commonly classified on the basis of the cement which binds the grains. Thus, there are *siliceous sandstones*, *calcareous sandstones*, and *ferruginous sandstones*.

Sandstones which are thoroughly silicified are known as *quartzites*. Although these rocks are classed as metamorphic, not sedimentary, they are mentioned at this point because all degrees of transition are found between sandstones and quartzites, and it is frequently difficult to tell where the rock should be placed. A practical basis for distinction is the fact that a quartzite is a very firm, compact rock which, on being broken, shows a fracture passing through the grains and the cement alike. A fracture in sandstone takes place in the cement, leaving the quartz grains unbroken.

Sandstones differ widely because of (1) differences in the size and chemical composition of the individual grains, and (2) differences in the cementing material. Consequently, it is difficult to generalize concerning the soils produced by sandstones as a group. It may be stated with certainty, however, that these soils will be rather coarse-textured. In addition, the more siliceous is the sandstone, the poorer will be the soil. On the other hand, calcareous sandstones generally give rise to good soils. This is to be expected, for the calcareous cement usually carries impurities, and the calcium carbonate itself is an important factor in determining soil reaction and general productivity. Wiedemann (1934: 36) noted that in Germany soils derived from sandstones rich in lime favor more abundant reproduction of spruce than do

soils developed from sandstones poor in lime. Additional observations by Hilgard (1906: 56), Schuchert (1930: 264), and Steven (1938) support the general view that highly siliceous sandstones and quartzites usually produce poorer soils than sandstones having a high content of lime or feldspars (arkose).

SHALE

Shale is formed as a result of consolidation of the finest particles of sedimentation. It is a close-grained and laminated rock. The rocks falling under this general designation are extremely variable in chemical composition. In contrast to sandstones, shales have very low porosity. Under the influence of pressure, shale becomes more and more compact and passes into a metamorphic rock known as *slate*. Because of metamorphic changes various minerals, for example, micas and chlorite, are formed and may appear as rather conspicuous elements.

Shales are to be regarded as relatively soft rocks, yet as a rule they produce shallow soils when they weather in place. Mechanical or physical weathering of both shales and slates proceeds more rapidly than does chemical weathering (Niklas, 1930: 89). The laminated structure and the dip of shales and slates have considerable influence on the rate of weathering. Furthermore, structure and dip are important from the standpoint of water relations and tree-root penetration; these influences will be discussed in the section dealing with schists. Soils derived from shales and slates are quite variable in quality; the end product of weathering is usually a heavy clay.

In various sections of the West, where annual precipitation is low, it has been noted that areas underlain with shale commonly support only grass or shrubby vegetation, whereas adjacent areas underlain with sandstone or igneous rocks support forest. This condition seems to be related to the more favorable moisture conditions in soils derived from sandstone or igneous rocks.

LIMESTONE

Limestone is a fine-grained rock consisting chiefly of calcium carbonate in more or less impure form. It effervesces freely in any common acid. In general, limestone is of organic origin, although it is also known to occur as a chemical precipitation;

chalk is probably derived from marine ooze. Igneous rocks are the primary source of the calcium of limestones. Limestones may alter to the metamorphic rock known as *marble*, which is distinguished from limestones by its crystalline nature, coarser grain, and greater compactness.

In general, limestone weathers rapidly, particularly the porous and impure varieties. The nature of the soil which develops from limestone depends to a considerable extent on the amount and kind of impurities contained, since the principal constituent, calcium carbonate, is dissolved and removed. In view of the great differences among limestones in kinds and amounts of impurities, it is not surprising that the derived soils are also extremely variable in quality. Soils derived from limestone are probably more variable than those derived from any other kind of rock (Ramann, 1905: 100; Domes, 1938: 473).

Soils derived from pure limestone (composed of calcite) may be exceedingly poor; they are frequently characterized by a low content of fine earth, by stoniness, and by dryness. In parts of Europe in which chalk, a rather pure form of limestone, occurs, the soils are notoriously poor (Hilgard, 1892; Domes, 1938: 473). Jolyet (1916: 337) remarked on the shallowness of certain soils derived from limestone in France. However, if more or less vertical fissures are present, tree roots may penetrate deeply and thus obtain moisture. Jolyet reported instances of calcareous oolites with foliations arranged more or less horizontally; such horizontal fissures are of little value for plant roots, and soils derived from such material may be very poor. It has been observed that very porous limestones may support plant growth, even in the absence of appreciable disintegration; nonporous limestones, on the other hand, support scanty vegetation unless disintegrated or decomposed (Kelley, 1934).

Moisture relationships in soils derived from limestones are sometimes unfavorable because of rapid drainage of water down cracks and cavities caused by solution of the calcium carbonate, particularly in shallow soils (Dengler, 1935: 412). In limestone areas the ground-water level commonly lies at considerable depths below the surface. However, soils derived from limestones should not, in general, be regarded as poor. On the contrary, in humid regions many of these soils are highly fertile; in fact, the remark has frequently been made, "A limestone country

is a rich country." Limestones containing fair proportions of clay materials weather to deep clay or clay loam soils, which, as a rule, have good underdrainage because of the openings in the underlying rock.

It should be pointed out that it is erroneous to use the term *calcareous soils* to designate all soils derived from limestone. Some such soils are, in fact, not calcareous at all, but highly acid. This statement is particularly true when chemical weathering of impure limestones has reached an advanced stage. On the other hand, young soils, in which physical weathering has been active, generally contain considerable amounts of calcium carbonate.

Limestone is known to exert an important influence on the development of soils and on the distribution and growth of forest trees. Podhorsky (1927: 394) cited Hess as showing that the upper altitudinal limit of beech stands in the European forest district of Oberhalsi is higher (1570 meters) on soils derived from calcareous rocks than on soils derived from primitive rocks (1330 meters). It has also been reported that, in the Dolomites and Carnic Alps, beech attains its highest limits on limestone and its lowest limits on siliceous rocks, regardless of exposure (Peattie, 1936: 94). In the interior of Alaska, white spruce (*Picea glauca*) extends up to 3000-ft. elevation on limestones, whereas in the same section the timberline on other formations is 2500 ft. or less.

It frequently has been observed that the eastern red cedar (*Juniperus virginiana*) is more abundant on limestone than on other formations. Examples of this relationship are known in Georgia, Missouri, Tennessee, Kentucky, and Alabama. A striking instance occurs in the northwestern part of Perry County, Missouri, where a bed of limestone (Everton) about 40 ft. thick lies between thick sandstone formations (St. Peter above, Everton below). A band of cedar trees on the limestone stratum can be clearly seen from a distance of 2 miles. In middle Tennessee there occurs an impure limestone called the Lebanon, named for the large cedar trees of the region. In the Southwest, Rocky Mountain juniper (*Juniperus scopulorum*) appears to be rather closely associated with limestone, and in the Ely district of Nevada, juniper is reported to be more abundant on limestone than on sandstone.

Soils derived from Permian limestones produce the heaviest and best stands of ponderosa pine in the Southwest, providing

precipitation is favorable in amount and distribution. In contrast to limestones the Coconino sandstones produce both poorer soils and poorer timber.² Kittredge (1928) reported that the best growth of jack pine (*Pinus banksiana*) in the Lake States was found on soil types where a limestone formation was present or where limestone had been brought in by glacial action.

Investigations in hardwood forest stands on soils derived from calcareous shales and sandstones northwest of Ithaca, New York, have demonstrated the occurrence of more of the exacting species, such as basswood (*Tilia americana*), white ash (*Fraxinus americana*), tulip tree (*Liriodendron tulipifera*), and hickory, than are found in stands on the acid soils to the south. The stands on soils derived from acid sedimentary rocks contain a higher percentage of American beech (*Fagus grandifolia*), yellow birch (*Betula lutea*), sugar maple (*Acer saccharum*), red maple (*Acer rubrum*), and certain oaks. These studies were made within a single climatic region so that the parent materials of the soils were the dominant variables.

DOLOMITE

Dolomite is a double carbonate of calcium and magnesium. In addition to dolomites, there are magnesian limestones showing considerable variation in composition. The term *dolomite* has been loosely used and is commonly applied to any limestone containing considerable magnesium. Like limestones, dolomite and magnesian limestones commonly contain impurities of various kinds. In the field, distinction between dolomite and limestone is usually impossible by inspection alone. Perhaps the most useful test is the reaction when treated with a cold dilute acid. Fragments of dolomite will not show active effervescence, but limestone will.

Dolomite alters less readily than limestone, with the result that stony soils of low productivity commonly develop. Hilgard (1906: 42) stated, "In some portions of Europe dolomite areas are sandy deserts, whose sand consists of weathered dolomite, so pure as to offer no adequate supply of mineral food to plants. In the United States, magnesian limestones underlie the 'barrens' "

² Personal communication from Quincy Randles, United States Forest Service.

of several states and thus seem to justify their European reputation of being poor soil-formers."³ On the other hand, very impure dolomites may give rise to good soils (Schucht, 1930: 265).

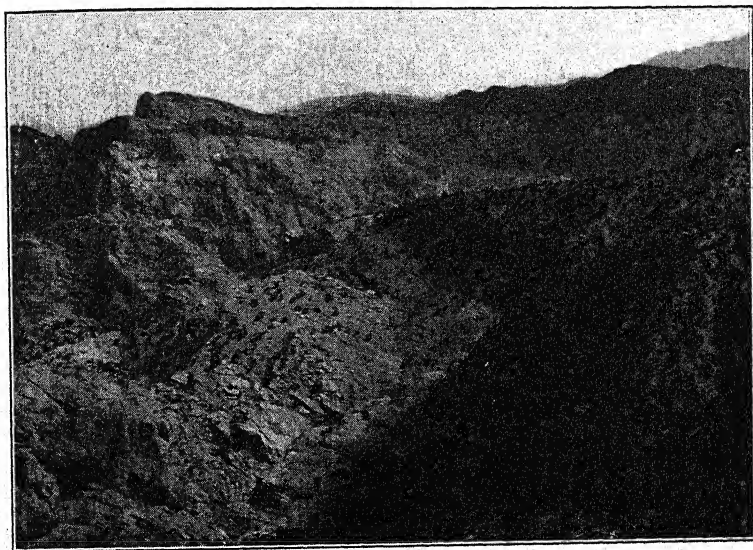


FIG. 7. The area on the right supporting piñon pine and juniper is largely dolomite; on the left the rocks are siliceous sandstones (southern Nevada).

Photograph by C. R. Longwell.

Figure 7 illustrates the pronounced difference in vegetation which occasionally may be noted at the contact of two dissimilar geological formations.

METAMORPHIC ROCKS

MICA SCHIST

A large variety of schists are recognized, but *mica schist*, which is essentially a mixture of quartz and mica with subordinate minerals, is the most common and important. Micaceous schists have a pronounced fissile nature due to the fact that the micas lie with their cleavage planes in the direction of schistosity.

Soils derived from mica schists resemble those developed from

³ From *Soils, their formation, properties, composition, and relations to climate and plant growth in the humid and arid regions*, by E. W. Hilgard. By permission of The Macmillan Company, publishers.

granite in having a low content of calcium and being coarse-textured. The tendency for schists to produce rather poor soils has been noted by Bowman (1911: 252), Jolyet (1916: 337), Schucht (1930: 261-262), and Aarnio (1934).

When standing more or less vertically, schists weather more rapidly than when horizontal, because penetration of water into planes of schistosity is favored. This fact was perhaps first observed by Grebe (1886: 77) and has subsequently been confirmed by Krauss *et al.* (1935) and others. Thin soils overlying schist rock which is standing upright are apt to be excessively drained and hence dry; when the soil body is fairly deep, the influence may be favorable because of the good underdrainage. When lying horizontally, schistose rocks in the soil or subsoil exert an unfavorable influence on both root penetration and the downward movement of water. In Saxony it was found that interruptions in the growth of Norway spruce (*Picea abies*) due to drought years were more serious in areas where schistose rocks were lying horizontally than in situations where the foliations were more nearly vertical (Wiedemann, 1925: 83). It is to be expected that trees growing on soils overlying schistose rocks having a vertical dip will be more wind-firm than those growing on soils overlying rocks in which the foliations are horizontal.

GNEISS

Gneisses form the largest group of metamorphic rocks and may be derived from either igneous or sedimentary formations. In common usage the term *gneiss* is practically restricted to granular, laminated rocks analogous in composition to granite, syenite, or diorite. Gneisses derived from igneous formations differ from them very little chemically. However, gneisses derived from sedimentary rocks are apt to be relatively poor in alkalis.

The rate of weathering of gneisses varies with differences in structure and mineral composition. Gneisses in which the laminations stand vertically, or nearly so, weather more rapidly than do similar rocks in which the laminations are nearly horizontal. Coarse-grained varieties weather more rapidly than fine-grained varieties, other factors being equal. Likewise, a relatively high content of feldspars and biotite favors more rapid weathering than does a high content of quartz and muscovite (Schucht,

1930: 259). In cool, humid regions mor humus layers commonly develop on soils derived from gneiss low in plagioclase feldspars and ferromagnesian minerals. General productivity of these soils may be expected to approximate that of soils from rocks of the rhyolite-granite clan.

SOME GENERAL RELATIONSHIPS BETWEEN ROCKS AND SOILS

In the preceding discussion some of the physical and chemical characteristics and also the general productivity level which may obtain in soils derived from certain rocks and rock groups have been indicated. Further consideration of this subject is necessary here in order that unsound interpretations may be avoided. The student should clearly understand that a given kind of rock will not everywhere weather to identical, or even similar, soils.

The most important factor disturbing any general relationship between rocks and the character and productivity of the derived soils is climate. It is now well known that soils developed from granite in New Hampshire are entirely different morphologically and chemically from those derived from granite in Georgia. In other words, pronounced differences in climate result in pronounced differences in soil, even when the parent materials are identical. This fact was not generally appreciated until the beginning of the twentieth century.

Early soil investigators in America and abroad leaned strongly toward a geological definition of soils. Evidence of this fact is the statement of Whitney (1892): "Geology defines the limits and areas of these different formations and of these different rocks, and as I have shown that these rocks determine the texture of the soil, a thorough and detailed geological map of the State [Maryland] should answer for a soil map." Smith and McCalley (1904: 74) also suggested that under certain circumstances a geological map might serve fairly well as a soil map. In succeeding years, however, when soil surveys were extended to more widely separated (and climatically different) geographical regions, the correlation between soil characteristics and geological characteristics was found to break down. The soil surveyor was at first much disturbed to find that soil derived from a given parent material in one region might be entirely different from that de-

rived from identical or similar parent material in another region. In discussing this situation Marbut (1928) pointed out that the difficulty of soil surveyors in correlating soil characteristics and geological characteristics arose principally when the attempted correlation involved soils in *widely separated areas*; it is significant that Marbut recognized that this difficulty "... affected only slightly the status of any given project considered alone." Thus the views of Whitney (1892) and Smith and McCalley (1904), applied to relatively small geographic units, were not groundless.

In 1914 there appeared the late K. D. Glinka's *Die Typen der Bodenbildung, ihre Klassifikation und geographische Verbreitung*, which made available the results of the Russian investigations concerning the role of climate in soil formation. There followed a somewhat general flight from the older geological basis and, as Marbut aptly put it, "a stampede to the Russian point of view." Geology was forsaken for climatology. Recognition of the importance of the new factor was at first accompanied, as it so often is, by neglect of the old.

The actual situation seems to be that the geographic occurrence of the broad zonal groups of soils is determined principally by climate and secondarily by vegetation. In relatively small regions of uniform climate, however, the nature of the parent material is probably more important than any other single factor in determining the character and productivity of soils.

Soil parent material and condition of the geological substrata are evidently of greater importance to the forester than to the agriculturist. In Chapter 2 it was pointed out that, as a soil body matures, its mineral composition departs more and more from that of the parent material. In other words, the influence of parent material on soil characteristics is most pronounced in young or immature soils and least in mature soils. This fact is of particular significance to the forester, since the soils with which he is concerned are commonly immature because of, among other reasons, occurrence on steep slopes where erosion may remove a substantial part of the products of weathering as they develop. Another point to be considered is the great depth to which tree roots penetrate, often extending far below the soil body into crevices in the underlying rock. In this connection it is of interest that C. A. Connaughton of the United States Forest

Service has observed roots of ponderosa pine in the Rocky Mountains extending to depths of 35-40 ft. in fractured granite. Presumably the advantage of this deep penetration of roots relates more to moisture supply than to mineral nutrients. Lastly, the forester is unable initially to select and subsequently to ameliorate his soils to the extent practiced by the agriculturist. The reason is the relatively low value of most forest crops.

Although there are many references in the literature to the relationship between surface geology and the distribution and growth of forest trees, detailed studies are less common. That such relations exist is indicated by numerous examples, several of which were earlier noted in the consideration of individual rocks or rock groups. The best examples usually occur on relatively immature soils. Northern white cedar (*Thuja occidentalis*) has been shown by Fernald (1919) to occur most abundantly on calcareous soils, and, conversely, the same authority has indicated that jack pine is most abundant on acid soils. It is significant that most of Fernald's observations were made in glaciated regions. Thus the influence of the parent rock is manifested even in soils derived from transported materials, such as glacial till. In New York state the authors have personally observed the common occurrence of northern white cedar on calcareous soils and the scarcity or absence of the species on adjacent acid soils.

A particularly striking instance of the influence of parent material on native vegetation is found in Alabama, Mississippi, and Texas (Hill, 1887, 1901; Hilgard, 1906: 53-54). In these states rendzina soils, developed from relatively soft calcareous beds, originally supported a natural vegetation consisting largely of prairie grasses and legumes, whereas adjacent red and yellow podzolic soils, derived from acidic parent materials, supported forest vegetation.

Tamm (1921), working in Sweden, recognized four groups of parent materials, based on calcium relations, ranging from acid quartzites with a low content of calcium to calcareous sedimentary rocks with a high content of calcium. Soils derived from the rocks lowest in calcium (quartzite, leptite) support a poor forest of Scotch pine. Rocks in the second group (granite, gneiss) are somewhat higher in calcium and produce good soils for pine and mixed conifers. The basic igneous rocks (diorite, gabbro, diabase, basalt) in the third group, together with the calcareous

sedimentary rocks in the fourth, have the highest calcium content and give rise to very productive soils which support herb-rich Norway spruce forests or, in southern Sweden, good hardwood forests.

The influence of the petrological nature of the parent material on general soil characteristics has been indicated a number of times in this chapter. Further consideration will be given to this important subject in Chapter 11, where soil development is discussed.

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Chapter 4

DISINTEGRATION AND DECOMPOSITION OF MINERALS AND ROCKS

The soil-forming minerals and rocks having been considered, it is now necessary to inquire how they are disintegrated and decomposed into soil material. All the minerals and rocks mentioned in the preceding sections may be present in the zone of weathering.

The agents of weathering are numerous, and the intensity of their action is variable within both space and time. To a large extent the weathering of minerals and rocks is controlled directly or indirectly by climatic conditions. This fact led Hilgard (1892) to suggest that there must exist a more or less intimate relation between the soils of a region and the climatic conditions which there prevail or have prevailed. The natural flora and fauna, which in large measure are a function of climate, also exert important influences. It may be readily appreciated that the operation of the agents of weathering is exceedingly complex.

For purposes of discussion it is convenient to distinguish between disintegration (physical weathering) and decomposition (chemical weathering). It should be borne in mind continually, however, that the two phenomena are closely associated and operate concurrently. It is obvious that disintegration favors decomposition, since disintegration results in greatly increasing the surface area exposed to chemical alteration. Likewise, decomposition favors disintegration, since a rock is weakened when its substance is chemically altered and in part dissolved. The mechanical and chemical forces act and react upon each other, each increasing the effectiveness of the other.

DISINTEGRATION OR PHYSICAL WEATHERING

In this type of weathering no chemical changes occur, and no new compounds are formed; the mineral or rock is merely broken up into smaller fragments. The resulting material is similar in

composition to the parent mineral or rock. Physical weathering is most intense in high latitudes or at high altitudes and in desert regions, particularly deserts at considerable elevation above sea level. Van Hise (1904: 496) stated that the regions favorable to weathering by disintegration, as contrasted to decomposition, are those of aridity, high latitude, marked topographic relief, and scanty vegetation.

Disintegration is accomplished by the following agents: water, ice, wind, temperature variations, change from water to ice, and plants.

WATER, ICE, AND WIND AS AGENTS OF DISINTEGRATION

The efficiency of these three agents in mechanical disintegration of rocks and minerals is largely dependent upon their use of rock materials as abrasive tools; wind, water, and ice themselves are incapable of rapidly cutting rocks. Ice in the form of glaciers, however, has accomplished a tremendous amount of rock disintegration. The glaciated areas of the United States alone exceed 500,000 square miles.

VARIATIONS IN TEMPERATURE AS AN AGENT OF DISINTEGRATION

Changes in temperature result in rock disintegration by producing differential expansion and contraction. Salminen (1932) reported that the rapid weathering of the Finnish rapakivi granite is in large measure due to temperature changes. More recently Polynov (1937: 11) stated that the immediate cause of disintegration of solid rock is usually fluctuation of temperature.

During the day rocks and soils may be heated to high temperatures because of exposure to direct solar radiation, and at night they cool rapidly. The alternate heating and cooling set up stresses within the rocks as a result of varying coefficients of expansion and contraction of their component minerals. Emerson (1920) stated that coarse-textured granite will disintegrate more readily than fine-grained granite. Rocks consisting of a number of coarsely crystalline minerals do not expand or contract uniformly. Emerson also maintained that color of rocks is of importance from the standpoint of disintegration, since dark-colored rocks become more heated than light-colored ones. Not

only do minerals of different colors heat differently, but also the same mineral seldom expands or contracts equally in all its dimensions. As a consequence stresses may be set up which in time result in breakdown of the rock.

The work of Barton (1916), Blackwelder (1925), and more recently Griggs (1936) casts some doubt on the effectiveness of naturally occurring temperature changes in rock disintegration. Griggs subjected a coarse-grained granite to artificial changes in temperature from 32° to 142° C. at 15-minute intervals. If each cycle is assumed to be equivalent to 1 day, the rock was subjected to the equivalent of 244 years of diurnal change. No alteration of the rock was evident in photomicrographs made before and after treatment. Radiant heat and dry cooling were utilized in this experiment. The failure of rocks to disintegrate when subjected to temperature changes in the laboratory does not disprove the theory that temperature changes in nature, continued over long periods of time, are effective in causing disintegration of rocks. Small strains, repeatedly developed, may have a cumulative effect which cannot be measured by laboratory experiments of relatively short duration.

Blackwelder (1927) has pointed out that in forested mountains of the western United States fire seems to be important in causing the disruption of boulders and rock outcrops. The characteristic effect of fire upon a rock mass is to expand the outer part so suddenly that the tensile strength of the rock is exceeded radially, and fractures are formed more or less parallel to the surface. Curved wedges, plates, or scales break off and leave a more rounded core, which is attacked in turn by the next fire. Gisborne (1932) also emphasized the importance of forest fires in rock disintegration. He pointed out that fire has been of universal occurrence in the past throughout all forest types in the United States.

CHANGE FROM WATER TO ICE AS AN AGENT OF DISINTEGRATION

The maximum density of water occurs at approximately 4.0° C.; its volume increases as the temperature falls below this point. Lang (1926: 278) gave the volume occupied by 1 gram of water at 0° C. as 1.00013 cc.; 1 gram of ice at 0° C. occupies 1.09083 cc.

Thus, when water changes to ice, it expands about 9 per cent. In general, rocks are not strong enough to resist the change of water to ice. The likelihood of disruption from this cause increases as the amount of contained water in the rock approaches saturation. Furthermore, the likelihood of disruption increases as the possibility of ready escape of water from the rock decreases. The maximum pressure which can develop when water freezes is close to 30,000 lb. per sq. in.; this maximum assumes certain special conditions, however, and it is highly improbable that it is ever attained in nature. Since conditions vary so greatly, it is practically impossible to specify just what pressures will develop when ice forms in rocks.

It has frequently been observed that rocks in the soil are heaved as a result of repeated freezing and thawing (Grawe, 1936). In this way they are brought into positions on or near the surface, where the agents of disintegration are most active.

PLANTS AS AGENTS OF DISINTEGRATION

It is well known that, as the roots of plants, particularly trees, develop in the soil body or in cracks in rocks, they exert considerable pressure and thus contribute to physical weathering. Lang (1926: 280) calculated that a root 10 cm. in diameter and 1 meter in length can raise a mass having a weight of 30,000–50,000 kgm. Even lichens, mosses, and similar forms of plant life are doubtless of significance in the disintegration of bare rocks. For example, Fry (1927) reported that swelling gelatinous apothecial or thallial tissue of lichens exerts an appreciable pull on the substratum to which it is attached.

Tree and other roots in the soil body thrust the loose material aside as they grow; and when they decay, they leave channels which favor easy penetration of water and permit surface material to fill in. The action of tree roots in splitting rock masses has been observed frequently (Cowles, 1901). Torrey (1925: 24) cited an instance in which the roots of a birch tree entered a crevice and lifted a boulder estimated to weigh 20 tons. At quarrying operations in Germany, pine roots have been found in fissures in sandstone at depths of 21 ft. (Vater, 1927). In the Rocky Mountains, as has been mentioned, ponderosa pine roots have been observed in fractured granite approximately 35–40 ft. below the surface.

When trees are windthrown, the soil mass is thoroughly agitated. Soil material from deeper layers, together with rocks of considerable size, is brought to the surface. This process may be quite important in soils having hardpan or incipient hardpan, since the material raised to the surface is then subjected to weathering. As the roots are torn out of crevices or fissures in the underlying rock, this substratum is frequently shattered. The significance of windthrow from the standpoint of weathering has been discussed by Van Hise (1904: 466) and Lutz (1940).

DECOMPOSITION OR CHEMICAL WEATHERING

In regions having fairly high humidity, low latitude, moderate topographic relief, and an abundance of plants and animals, decomposition is favored more than disintegration. Decomposition leads to profound changes in minerals. Certain of them may disappear either wholly or partially and be replaced by entirely new minerals, which are formed *in situ* or precipitated from solution. Soils resulting from intense chemical weathering are, as a rule, quite different in composition from the original mineral or rock material. In general it may be said that intense chemical weathering causes soils to be more alike in chemical composition than are the rocks from which they were derived. Chemical weathering is accomplished by plants, animals, water, and gaseous solutions.

PLANTS AS AGENTS OF DECOMPOSITION

The organic complex of the soil, which is derived chiefly from plant remains, contains both inorganic and organic acids. As examples of the inorganic acids, sulfuric, nitric, and phosphoric acid may be mentioned. Organic acids, such as formic, acetic, propionic, malic, oxalic, and levulinic, also occur. The inorganic acids appear more effective in mineral and rock weathering than the much weaker organic acids (Blanck and Geilmann, 1924; Behrend, 1931; Blanck, 1933).

Mineral weathering appears to proceed more rapidly in the vicinity of tree roots than at points where they are lacking (Van Hise, 1904; Clarke, 1924b: 487; Metzger, 1928; Waksman and Starkey, 1931: 161). It has been observed that the soil or decomposed rock around tree roots is often bleached by solution

and removal of the iron. Formation of clay pockets in the English chalk has been associated with the presence of roots and their by-products (Anderson, 1927). The higher concentration of carbonic acid in the vicinity of roots undoubtedly assists in chemical decomposition of the soil minerals and may be a favorable influence from the standpoint of plant nutrition. Metzger (1928) found the concentration of bicarbonates higher in soil close to roots than at some distance away.

Bacteria and soil fungi also contribute to weathering. Various products, such as carbon dioxide and organic and inorganic acids, are produced by the metabolism of soil microorganisms. These products aid in mineral decomposition, especially carbonates and phosphates. Under conditions favoring anaerobiosis, certain compounds rich in oxygen, such as nitrates and sulfates, may be used by bacteria as sources of oxygen. Mineral elements, such as iron, sulfur, and manganese, or their inorganic compounds may also be used by certain bacteria as sources of energy. The importance of microorganisms in the weathering of rocks and soil material is widely recognized. According to Van Hise (1904: 505), living organisms are an essential factor in much of the chemical work of decomposition and solution of rocks.

The influence of vegetation on chemical weathering of the soil material varies with differences in floristic composition. Chemical weathering is most intense under vegetation having a low base content. Jolyet (1916: 324) stated that chemical decomposition is more active in soils supporting forest stands than in cultivated soils. The difference may be explained by the influence of the tree roots and the organic matter of the upper soil layers.

ANIMALS AS AGENTS OF DECOMPOSITION

Animals, like plants, accomplish chemical work both while alive and after death. The decomposition of animal remains on and in the soil results in the production of carbon dioxide, nitrates, and other chemically active substances which assist in weathering.

The influence of vertebrate animals on chemical decomposition of soil material is poorly understood. However, work by Greene and Reynard (1932) and Greene and Murphy (1932) in the

desert soils of Arizona is suggestive. Notable increases in calcium, magnesium, bicarbonate, and nitrate ions were found in the dens of the kangaroo rat and the pack rat.

Various insects, particularly ants and termites, undoubtedly contribute to chemical alteration of the soil material (Mills, 1889). They carry organic matter into the soil, where it decomposes with the formation of products that favor weathering of the mineral matter. Likewise, the soil body is rendered more permeable to both air and water by their tunnels and galleries.

Earthworms perform particularly valuable services in the soil. Darwin (1882) believed that their principal work was the separation of the finer from the coarser particles, mixing the mineral constituents with organic matter, and saturating the mixture with their intestinal secretions. It has been suggested by Van Hise (1904) that repeated passage of soil material through the bodies of earthworms may result in dissolving very appreciable amounts of minerals and in rendering undissolved constituents more soluble. Myriapods may also carry on effective work (Romell, 1935).

WATER AND GASEOUS SOLUTIONS AS AGENTS OF DECOMPOSITION

Ground water is very different from rain water. When rain water penetrates deeper and deeper into the ground, it undergoes progressive modification in composition. Ground-water solutions usually contain salts derived from the acids and bases of rocks; such common elements as sodium, potassium, calcium, and magnesium may exist as carbonates, sulfates, and chlorides. In addition, the ground water contains oxygen, carbon dioxide, and other gases in solution. Of the dissolved gases probably oxygen and carbon dioxide are the most important.

Water is of such profound importance from the standpoint of rock alteration and soil development that it has been compared with the blood of an organism (Van Hise, 1904: 63; Clarke, 1924a). Chemical weathering is closely associated with water; in very dry or very cold regions chemical weathering is slight. Since practically all compounds are attacked by water, it is sometimes referred to as the universal solvent. The presence of carbon dioxide in water adds materially to its activity. Under the influence of carbonated waters the pyroxenes and amphiboles

alter most readily, followed by the plagioclase feldspars; orthoclase and the micas are more resistant (Clarke, 1924b).

Much of the material which goes into solution as water passes through the soil is removed in the drainage water. In humid regions the dominant salts carried by rivers are carbonates; in arid regions sulfates and chlorides prevail (Clarke, 1924a). The average composition of the dissolved matter in North American fresh waters has been calculated by Clarke with the results shown in Table 5. These values are similar to those for the

TABLE 5. AVERAGE COMPOSITION OF NORTH AMERICAN WATERS

(After Clarke, 1924a)

Element or Compound	Per Cent
CO ₃	33.40
SO ₄	15.31
Cl	7.44
NO ₃	1.15
Ca	19.36
Mg	4.87
Na	7.46
K	1.77
(Fe,Al) ₂ O ₃	0.64
SiO ₂	8.60
	<hr/> 100.00

Mississippi River alone. The relative mobility of certain constituents has been indicated by Polynov (1937: 163) as follows: Cl > SO₄ > Ca > Na > Mg > K > SiO₂ > Fe₂O₃ > Al₂O₃. The order of solubility given by Schuchert (1930: 91) is only slightly different: chlorides > nitrates > sulfates > carbonates of calcium and magnesium > phosphates.

Temperature relations are important, since they influence the activity of solutions in the rocks and soil; as a general rule, the higher the temperature, the more rapid the alteration. Particularly important from the standpoint of chemical weathering is the length of time during the year when temperatures are above 0°C. (Obst *et al.*, 1930: 145). Going from high to low latitudes involves passing from regions of minimum to regions of maximum chemical weathering. In the tropics the zone of weathering may extend as deeply as 600 ft. (Blanch, 1929: 224). Russell

(1889) reported that crystalline rocks, such as granites and schists near Raleigh, North Carolina, are in places weathered to a soft, plastic clay at depths exceeding 50 ft. Similar observations have been made by Whitney (1928), who stated that the igneous rocks of the Piedmont Plateau are rather well decayed to a depth of 40 ft.; 20 ft. down, the rock material is so decomposed that it has taken on certain characteristics of soil.

JOINT WORK OF THE AGENTS OF DECOMPOSITION

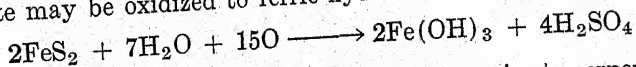
It should be recognized that the agents of weathering commonly work together. Numerous reactions of weathering are the result of joint action, and it is convenient to consider them in this light.

OXIDATION

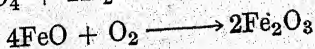
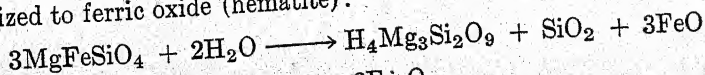
Oxidation involves a loss of electrons or negative charges. In many reactions it consists of the addition of oxygen, but in some oxidation reactions no oxygen changes occur. The source of oxygen is the atmosphere, which contains 23.12 per cent of the gas by weight. The atmospheric oxygen acts to some extent directly, but to a far greater extent through water solution, and to the greatest extent through water solutions and organisms combined. Of these organisms bacteria are of the greatest significance, but molds and fungi are also important.

As a result of oxidation of plant debris, various organic acids are formed as intermediate products. The end products of oxidation of carbohydrate compounds are carbon dioxide and water. Carbon dioxide, as has been noted, is highly effective from the standpoint of weathering. Nitrogen in the complex protein compounds is oxidized to nitrites and to nitrates.

One of the most important inorganic compounds to be oxidized is iron. When this element occurs in the carbonate, sulfide, or silicate form, it is particularly susceptible to alteration. Iron pyrite may be oxidized to ferric hydroxide and sulfuric acid:



As a result of hydrous alteration, olivine gives rise to serpentine, silica, and ferrous oxide. The ferrous oxide is subsequently oxidized to ferric oxide (hematite):



Magnetite may be oxidized to hematite, and simultaneously hydration may occur, giving rise to limonite or other hydrated oxides. If iron carbonates are oxidized in the absence of hydration, hematite is formed; with hydration, limonite forms. Oxidation of the ferrous oxide of silicates to ferric oxide (either hydrous or anhydrous) occurs simultaneously with processes of carbonation and hydration of the other bases. It may be pointed out that volume changes arise from oxidation. Where no subtraction of material occurs, as may be the case in oxidation of the ferrous iron of silicates, the process results in an increase of volume; these volume changes favor further weathering.

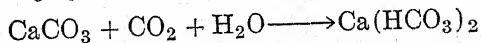
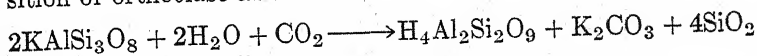
Although oxidation may properly be regarded as the normal process in the zone of weathering, under certain conditions reduction occurs. Reduction can be looked upon as the reverse of oxidation, that is, the gain of electrons, and both processes may go on in the soil at the same time. Reduction processes are most strongly developed in wet situations where organic matter is present and aeration is deficient, for example, bogs, marshes, swamps, and other poorly drained sites. Bacteria acting on the organic matter obtain their oxygen from compounds, such as ferric oxide and ferric hydroxide, in the higher states of oxidation. As a result, ferric oxide, for example, is reduced to ferrous oxide, sulfates are reduced to sulfides, and nitrates to nitrites or ammonia. In soils where reduction processes have been active, gray or other light colors usually predominate. Reduction tends, particularly in iron and manganese compounds, to increase mobility.

CARBONATION

Carbonation is a process involving the combination of carbonic acid with bases to form carbonates. Plants absorb carbon dioxide from the atmosphere, where the concentration is relatively low, and utilize it in the synthesis of organic compounds. Upon death of the plants their tissues are decomposed, and the carbon dioxide is again liberated. As a result a relatively large amount of carbon dioxide is concentrated in the zone of weathering and is available for carbonation. Carbonation is of fundamental importance from the standpoint of mineral and rock alteration. In cold regions it proceeds slowly but is active in warm, humid regions having luxuriant vegetation.

Because carbon dioxide is easily soluble, it occurs as carbonic acid in all natural waters. Rain water reaching the ground contains carbonic acid, and in seeping through the upper soil layers it is further fortified. Although carbonic acid is weak because of the low dissociation of its hydrogen ions, it is present in large quantities and in time may effect drastic changes.

The process of carbonation may be illustrated by the decomposition of orthoclase and calcite:



In the first reaction the products consist of kaolinite, potassium carbonate, and silica; in the second, calcium bicarbonate is formed. Van Hise (1904: 478) has pointed out that carbonation is frequently accompanied by oxidation and hydration; desilicification may or may not occur. In all the various combinations of reaction, provided the compounds formed remain as solids, there is an increase in volume; this increase commonly amounts to 15-50 per cent.

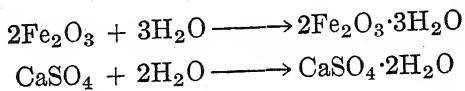
To a considerable extent carbonation operates through the decomposition of silicates; silica is set free as colloidal silicic acid. Carbonates are also formed as a result of combination of carbon dioxide with oxides, such as the ferrous oxide in magnetite. When alkaline carbonates have been generated in the ground water, its solvent power is greatly increased; such waters then become effective in the solution and redeposition of silica (Clarke, 1924b: 486).

The carbonates of potassium and sodium and the acid carbonate of calcium are all soluble in water and in humid regions tend to be carried out in the drainage waters; in arid regions the carbonates may accumulate as alkali.

HYDRATION AND DEHYDRATION

Hydration consists of the addition of water to chemical substances, with the formation of hydrous compounds. This process is very widespread in the zone of weathering and may be especially important in the early stages of rock alteration. As weathering proceeds, hydration increases. Among the hydrous minerals are many silicates and oxides and some carbonates and sulfates.

The process of hydration may be illustrated by the transformation of hematite to limonite and the change of anhydrite to gypsum:

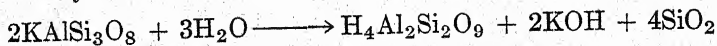


As a result of hydration, minerals may become softer and increase in solubility. Burt (1927) stated that the average hardness of nine common anhydrous soil minerals which are easily hydrated is 6.1. The average hardness of the six most common soil minerals formed by hydration of the foregoing minerals is 3.7. Hydration also leads to an increase in volume. The alteration of anhydrite to form gypsum is accompanied by an increase of volume amounting to about one-third (Rastall, 1916: 44). Such volume increase favors disintegration.

Although hydration processes are the rule in the zone of weathering, dehydration may also occur. Probably dehydration operates most extensively in regions of high temperature where wet and dry seasons alternate, that is, in some of the tropical regions. During the wet periods hydration occurs, and during the dry periods there is a reversal, with dehydration becoming operative. Ferric iron in the belt of weathering is usually hydrous, giving rise to yellow or brown soil colors. However, in regions of high temperature where the humidity is low for at least a part of the year the soil may be red because the iron is present in anhydrous form as hematite (Van Hise, 1904: 482). Dehydration is most marked in the surface soil.

HYDROLYSIS AND SOLUTION

Hydrolysis and solution are closely related and may be considered together. Hydrolysis, which is highly important in the chemical weathering of rocks, consists essentially of the double decomposition of water and a salt. The reaction may be illustrated by the alteration of orthoclase to kaolinite:



The concept of hydrolysis is based on the fact that water dissociates into H^+ ions and OH^- ions and that the H^+ ions are capable of replacing other cations, such as K, Mg, Na, and Ca. The process is hastened by the presence of inorganic and organic acids

in the soil solution. High temperatures favor intense hydrolysis, whereas in cool climates this process is relatively weak (Harrassowitz, 1930: 362; Blanck and Giesecke, 1933).

The general results of hydrolysis may be outlined as follows:

1. *Desilicification occurs.* During the weathering process in humid regions there is generally a loss of silicic acid, which is removed by percolating waters.

2. *Dealkalization occurs.* In humid regions weathering is accompanied by losses of calcium, magnesium, potassium, and sodium.

3. *New compounds are formed.* These compounds are referred to as the clay complex or the inorganic colloidal complex.

As the various processes of weathering operate, compounds pass into solution in the ground water. Both carbonation and hydration result in compounds of higher solubility. The change of ferrous to ferric iron by oxidation has a reverse effect, but the quantitative value of this reaction is small compared with that of carbonation and hydration (Van Hise, 1904: 484). As already noted, all naturally occurring compounds are soluble in pure water to varying degrees. However, the soil solution and underground waters are not pure but contain salts and various acids which appreciably increase activity.

Materials are continuously being removed in solution from the zone of weathering. The residues left behind are, of course, the least soluble constituents, and they form the soil. All the elements in the soil material are being dissolved but at greatly varying rates. Elements such as sodium, potassium, calcium, and magnesium are more easily removed in solution than are iron, silicon, and aluminum.

GENERAL RESULTS OF ROCK WEATHERING

The products resulting from rock weathering vary with the kind of rock, the environmental conditions under which weathering occurs, and the extent of the alteration. For this reason it is unwise to generalize broadly as to the nature of the products resulting from weathering of different rock types. Nevertheless it seems desirable to point out some of the major changes which occur.

After consideration of the analyses of rocks and their decomposition products Clarke (1924b: 492) pointed out that in nearly all

weathering certain features seemed clear. Thus the iron compounds are oxidized, and there is a notable assumption of water by the undissolved residues. The soluble bases are lost by leaching, and alumina and ferric oxide are concentrated in the residual materials. Much of the dissolved silica is carried away.

In Table 6 are presented computations by Merrill (1921) which show some of the results of rock weathering.¹ The computations are based on analyses of rocks and their decomposition products as they occurred at a number of points in the eastern United States.

TABLE 6. RESULTS OF ROCK WEATHERING

(After Merrill, 1921)

	Granite	Gneiss	Diabase	Diorite
<i>A. Percentage Loss for Entire Rock</i>				
SiO ₂	10.50	31.90	33.41	17.43
Al ₂ O ₃	0.46	Standard	9.11	Standard
FeO, Fe ₂ O ₃	Standard	1.30	Standard	3.53
MgO	0.36	0.80	15.09	4.97
CaO	0.81	4.44	9.71	9.20
Na ₂ O	0.77	2.68	2.64	2.17
K ₂ O	0.85	3.55	0.35	0.21
P ₂ O ₅	0.04
	13.79	44.67	70.31	37.51
<i>B. Percentage of Each Constituent Lost</i>				
SiO ₂	14.89	52.45	73.64	37.31
Al ₂ O ₃	3.23	Standard	68.19	Standard
FeO, Fe ₂ O ₃	Standard	14.35	Standard	21.03
MgO	1.49	74.70	98.81	97.17
CaO	25.21	100.00	98.68	97.30
Na ₂ O	28.62	95.03	82.46	84.87
K ₂ O	31.98	83.52	77.31	38.75
P ₂ O ₅	40.00	19.87

Granite. A gray micaceous granite from the District of Columbia. Fresh rock and residual sand.

Gneiss. A biotite gneiss or gneissoid granite from near North Garden, Albemarle County, Virginia. Fresh rock and plastic, red-brown, residual soil.

Diabase. A dark gray, medium-textured olivine diabase from near Chatham, Virginia. Fresh rock and its bright orange, residual clay.

Diorite. A fine-grained, almost coal-black diorite from North Garden, Albemarle County, Virginia. Fresh rock and its residual soil.

¹ From *A treatise on rocks, rock-weathering, and soils* by George P. Merrill. By permission of The Macmillan Company, publishers.

In setting forth the alterations which rocks have undergone as a result of weathering it is common practice to assume that one rock constituent is constant and then to compare the others with it. Either ferric oxide or alumina may be taken as constant, and with it as a standard the relative losses of the other constituents can be estimated.

Merrill (1921) used the following formula in his calculations:

$$\frac{A}{B \times C} = X$$
$$100 - X = Y$$

In this formula A = the percentage of any constituent in the residual material; B = the percentage of the same constituent in the fresh rock; and C = the quotient obtained by dividing the percentage amount of alumina (or iron sesquioxide, whichever is taken as a constant) of the residual material by the amount in the fresh rock, the final quotient being multiplied by 100. The expression X then = the percentage of the original constituent saved, and Y = the percentage of the same constituent lost.

In considering such data on rock weathering as are presented in Table 6, it should be borne in mind that what is indicated is the ultimate composition; the manner in which the elements are combined in the fresh rock and in the weathered product may be, and commonly is, quite different.

In the granite cited in Table 6 the weathered material is more a disintegration product than a decomposition product. In other words, it represents changes occurring in the earlier stages of weathering. The gneiss from Virginia shows more drastic alteration, having been weathered to residual clay. It is noteworthy that the alkalis and alkaline earths have suffered the greatest loss; 95.03 per cent of the soda, 83.52 per cent of the potash, 74.70 per cent of the magnesia, and all the lime have disappeared. In the diabase from Virginia we have another example of advanced weathering. Evidently there has been a total loss amounting to 70.31 per cent, or more than two-thirds of the original material. The diorite analysis also indicates intense weathering.

The data in Table 7 on the results of lateritic weathering of gneiss from Colombo have been calculated from the analyses of Polynov (1937). The content of Fe_2O_3 plus FeO is assumed to have remained constant.

DISINTEGRATION AND DECOMPOSITION

TABLE 7. RESULTS OF ROCK WEATHERING

	(After Polynov, 1937)			
	Fresh Gneiss, per cent	Weathered Material, per cent	Percentage of the Original Constituents	
			<i>Saved</i>	<i>Lost</i>
SiO ₂	64.42	33.87	21.09	78.91
Al ₂ O ₃	17.25	26.57	61.80	38.20
Fe ₂ O ₃	4.21	19.69	100.00
FeO	3.95	0.65		
MgO	0.49	100.00
CaO	2.10	100.00
Na ₂ O	2.42	Trace	100.00
K ₂ O	2.99	Trace	100.00
Organic matter +H ₂ O	0.94	17.82	100.00

Here the weathering has been very intense, with complete loss of the magnesium, calcium, sodium, and potassium. The high loss of silica is also noteworthy.

Polynov (1937) recognized four phases in the process of weathering. In phase I the products of weathering lose their compounds of chlorine and sulfur. In phase II the products which have already lost their chlorine and sulfur compounds begin to lose the bases calcium, sodium, magnesium, and potassium. In phase III most of the bases have been removed, and the silica of the aluminosilicates becomes mobile. Finally, in phase IV the residual products consist largely of sesquioxides. Polynov recognized that climate has a substantial influence on the intensity of the weathering process but believed that the general direction remained unaltered.

THE PRODUCTS OF WEATHERING

Commonly the products of mineral and rock weathering (the parent material from which the soil profile develops) are divided into two great groups: (1) sedentary or residual, and (2) transported. The sedentary products are those derived from the weathering of solid rocks in place; the transported materials have been moved from the point where the parent rock outcropped, for example, glacial drift, river alluvium, or loess. In sedentary materials the origin may be a single lithologic unit; in transported materials the origin is commonly several different units. These facts are of significance in studies of the relationship of surface geology and soil character.

SEDENTARY MATERIALS

Residual soil material is not stratified and contains no matter foreign to the locality. Residual soils are most abundant in the portion of the United States east of the Mississippi River and south of the glaciated areas (Merrill, 1921). They are well known in the Piedmont region, the Appalachian region, and the great limestone valleys of Kentucky and Tennessee.

Residual soils vary greatly, being influenced by the regional climatic conditions, topographic situation, character of the parent rocks, nature of the vegetation supported, and age of the material. Chemical deficiencies are more apt to appear in soils derived from residual material than in those developed from transported material. A soil derived from transported material usually contains mineral substance from a variety of rocks and consequently is more apt to be balanced in nutrients than is a soil derived from only one kind of rock.

Merrill (1921) used the term *cumulose deposits* to indicate the accumulation of organic debris in swamps, lakes, or ponds. The accumulation is regarded as occurring *in situ* with only relatively small amounts of foreign detritus. As a rule, the proportion of organic matter in cumulose deposits increases from the margin to the center.

TRANSPORTED MATERIALS

The products of weathering commonly do not remain long *in situ* where formed. Forces such as water, wind, ice, and gravity are powerful agents of transportation. Frequently two or more agents of transportation have been active in the formation of a given deposit. The principal groups of transported materials may be classified on the basis of the agencies primarily responsible for their movement and deposition.

1. *Materials moved by water*
 - A. Alluvial deposits
 - B. Lacustrine deposits
 - C. Marine deposits
2. *Materials moved by wind*
 - A. Loess deposits
 - B. Dune sand deposits

3. *Materials moved by ice*

A. Glacial till deposits

4. *Materials moved by gravity*

A. Colluvial deposits

Alluvial deposits are formed by streams and may occur either as flood plain depositions or as alluvial fans. The competence (ability to transport in terms of dimensions of particles) of a stream varies as the sixth power of the "bed" velocity. Thus a stream whose velocity is 5 miles per hour can transport particles having a size or weight of X , and a stream whose velocity is 10 miles per hour can transport particles having a size or weight of $64X$. Deposition occurs whenever the velocity of the stream falls below certain critical levels, which vary for particles of different sizes. It is evident at once that alluvial sediments may vary tremendously. Rapidly moving streams transport and deposit coarse gravel and sands, whereas silt is deposited from sluggish waters.

As a rule, alluvial materials are more or less distinctly stratified, and cross-bedding is common on flood plains. Because of stratification in these deposits it is necessary to be cautious in accepting the underlying strata as the parent material from which the soil developed. Frequently the nature of the deeper-lying strata is of high importance to tree growth. For example, the strata may vary in mineral composition or water relations (von Kruedener, 1934: 118).

It has been pointed out by Emerson (1920) that alluvial material along large rivers, especially along the lower reaches, is less variable in texture than that along smaller streams. Along the lower reaches of a river the currents are less variable than along the upper parts. It may also be pointed out that flood plain deposits commonly show increasing coarseness of particles with increasing depth below the surface. Wide variability is also noted in the mineral composition of the alluvial sediments. Alluvial fans develop at the bases of slopes. They usually contain considerable coarse material and are well drained.

Alluvial materials, if they are well drained and not too coarse-textured, generally give rise to productive soils. Harrassowitz (1930: 367) has pointed out the prevalence of alluvial deposits

in tropical regions where erosion is considerable because of alternation of wet periods, with heavy rains, and of dry periods.

Lacustrine deposits are laid down in lake basins. The sediments vary greatly in texture, being coarse near the shores and finer farther out. Clay and silt soils, developed from the finest particles of sedimentation, are frequently of lacustrine origin. As might be expected, a relationship exists between the nature of the lacustrine material and that of the rocks which form the drainage basin. This relationship is stronger in small lake basins than in large ones.

Finley (1937) and others have pointed out the influence of beaver in the development of valley meadows. The ponds created by the beaver served as settling basins and in time were filled with water-borne mineral matter and the remains of plants.

Marine deposits are formed in ocean basins. They commonly contain a high proportion of particles of quartz and other refractory minerals and may be low in nutrient elements. Stratification is common. In the Atlantic and Gulf coastal plains marine deposits are of general occurrence.

Loess deposits result from wind action. It is believed that much of the loessial material was deposited during Pleistocene time. At present loess is being deposited at a rapid rate in the Matanuska Valley in Alaska (Tuck, 1938). The material is derived from glacial silt on the flood plains of the Matanuska and Knik rivers. As is true of other types of deposit, it is frequently difficult to ascribe the origin of a formation to wind action alone, since water or other types of transport also may have occurred. It cannot be doubted, however, that wind is a powerful agent of transportation. In recent years, particularly, the dust storms in the western states have demonstrated the effectiveness of this agency in transporting soil material long distances.

Loess is remarkably uniform in mechanical composition. The silt content is generally high, with some clay and rather small amounts of sand. A characteristic feature of loess is the vertical or columnar face presented in the profile. Road cuts often become vertical-walled canyons as the material is eroded.

Sand dune deposits, like loess, result from wind activity. They are common features along the east side of Lake Michigan and at various points on the Atlantic and Pacific coasts. The dune sands are usually highly siliceous and produce poor soils as a

rule. Their lack of stability and poor moisture-holding capacity make them difficult sites for forest vegetation.

Glacial till results from glacial activity and includes the drift materials which are primarily a product of ice action. Actually their deposition may have been due in part to water. Materials of strictly aqueoglacial origin are classified as alluvial.

Glacial till is composed of rocks and their products of weather-

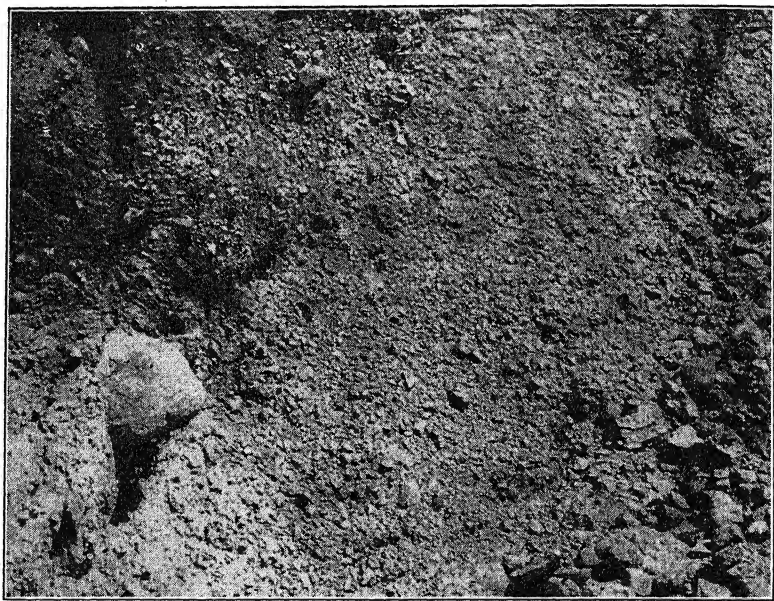


FIG. 8. Exposure of Illinoian till, Stony Island, Detroit River, showing the lack of size sorting. Photograph by United States Geological Survey.

ing. The material has been mechanically disintegrated to varying degrees. In contrast to water and wind, glacial ice transports all rock fragments, irrespective of size. There is no assortment of materials, as there is in water- and wind-laid deposits (Figures 8 and 9). Rocks of many different sources are commingled; some have travelled far, others very short distances. The mixtures vary tremendously and are governed only by the nature of the parent materials passed over by the ice. In places where the ice moved for some distance *along* the outcrop or strike of a formation, the drift is composed predominantly of material from

that outcrop. If, on the other hand, the ice moved *across* several rock outcrops, the mixture may be one of great complexity.

In general, glacial till soils are heavier-textured than those developed from alluvial materials. Occasionally the till is extremely compact and relatively impervious to water. Oskamp (1935) observed in New York that apple trees root more deeply in gravelly soils which were water-laid than in such soils derived

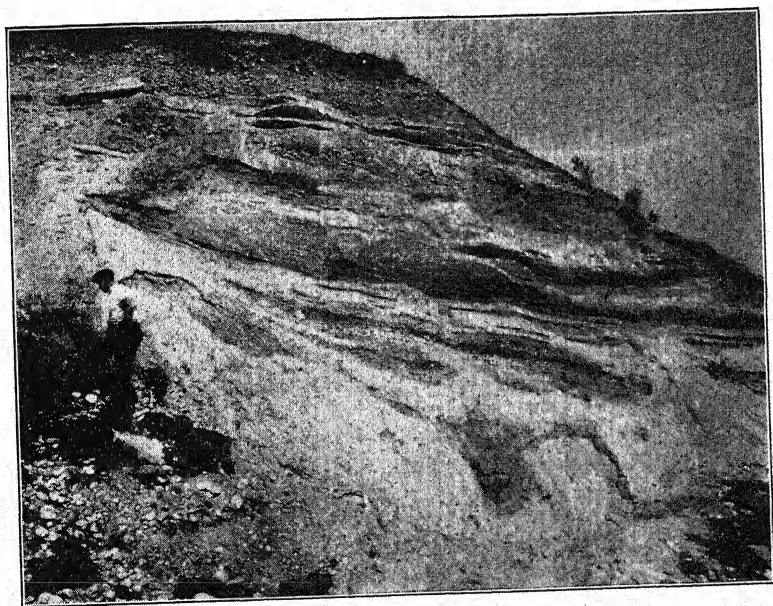


FIG. 9. Stratified deposits built up in a temporary glacial lake in the valley of Pawcatuck River, north of Pawcatuck, Connecticut. Photograph by R. F. Flint.

from glacial till. Frequently the till, particularly if laid down during the Wisconsin period of glaciation, is only slightly weathered; the resulting soils are usually immature. The importance of glaciation in soil development in North America can scarcely be exaggerated. The influence on soils has been tremendous both in this country and in much of northern Europe.

Colluvial deposits are the accumulations of rock detritus commonly recognized as talus and cliff debris; material transported in avalanches and landslides is also classified as colluvial. Gravity is the principal agent of transportation. These deposits are

usually of limited area and as a rule contain a large amount of rock. They generally occur on or at the base of rather steep slopes in regions of precipitous topography. The weathering is predominantly disintegration; hence there are commonly only small amounts of fine earth present between the rocks, principally in the lower part of the bed and at the lower limits of the slope. Shaler (1891) has described in detail the development of colluvial deposits.

The so-called soil creep or solifluction (Andersson, 1906; Högbom, 1914; Denny, 1936; and others) is very slow movement, almost particle by particle, of the products of weathering down slopes. The movement is due to gravity, and its rapidity is determined by the degree of slope. Emerson (1920) emphasized the importance of freezing and thawing in the process. As water in the soil freezes, the particles are lifted slightly; when the mass thaws, the particles sink vertically. On level ground the material returns to its former position, but on slopes it is displaced horizontally, thus moving downhill. Sheet wash or sheet erosion may be regarded as a special kind of soil creep.

Howe (1909) made a study of landslides in the San Juan Mountains, Colorado. He remarked that landslides are common in regions which are topographically young or just approaching maturity. Saturation of the ground by water was recognized by Howe as a condition favoring landslides. The most comprehensive recent study of landslides and other types of mass movement is contained in the monograph by Sharpe (1938). This valuable publication, which should be consulted by those particularly interested in the subject, contains an excellent review of the vast literature.

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Chapter 5

FOREST-SOIL ORGANISMS

The importance of plant and animal organisms in soil development and soil conditions can hardly be overemphasized. Vernadsky (1928) expressed the view that the chemical composition of the biosphere of the earth may be considered the result of the activities of living organisms. From the standpoint of forest vegetation, soil organisms are particularly important, as has been pointed out by Müller (1887), Bornebusch (1930), and others. Their significance is probably even greater in forest soils than in agricultural soils.

Large amounts of organic matter, as will be shown in Chapter 6, are deposited annually on the surface of forest soils. This material must be disintegrated and decomposed if the contained nutrient elements are again to become available for use by higher plants. For maximum benefit it is essential that the organic matter or its decomposition products become incorporated in the upper layers of mineral soil. The disintegration, decomposition, and incorporation of organic matter are highly important processes and are largely effected by the macroscopic and microscopic organisms which abound in all fertile forest soils. The species composition and abundance of the fauna and flora of soils vary with factors such as climate, physical and chemical properties of the soil, and type of vegetation. In turn, these organisms influence the type of humus layer, soil profile development, and physico-chemical properties of the soil. Thus an intimate relationship exists between the life in the soil and the forest stand which the land supports.

In this chapter a brief account is presented of the more important soil organisms. Consideration is given to the abundance and ecology of the various species or groups to the extent that such information is available. Actually much further research is necessary before a well-balanced and reasonably complete picture of the fauna and flora of forest soils can be drawn.

FAUNA OF FOREST SOILS

LAND VERTEBRATES

The influence of higher animals on soil operates principally in cultivation effects and the breakdown of the organic matter. Many animals burrow in the soil and over a period of years accomplish considerable work. Material from subsurface horizons is brought up; and surface material, both organic and inorganic, either is carried down or gradually works down by gravity. In this way the soil material is mixed and kept in a friable condition. Penetration of water and air into the soil body is greatly facilitated. Decomposition of organic debris is hastened by animal activity as a result of its use as food and as a result of its being mixed with the mineral soil material.

Animals such as the prairie dog, rabbit, mole, badger, woodchuck, gopher, mouse, shrew, and ground squirrel have been recognized as important in soil development by various authors. Particular emphasis has been placed on the work of moles by Müller (1887), Ramann (1911b: 493), Hauchecorne (1927), and Bornebusch (1930). Ramann regarded moles as the most important soil-working animals in the German forests, and both Müller and Bornebusch pointed out their wide occurrence in Danish forests having mull humus layers; they appear to be absent from many soils having mor types of humus layers. The tunnels made by moles facilitate entry of water and favor its distribution in the soil body. Hoffman (1931: 428) stated that moles are very widely distributed, occurring in the greater part of Europe and Africa and in North America. The habits of these animals have been investigated by Hisaw (1923a, 1923b) and Hamilton (1940), who found that they were predominantly carnivorous. They devour earthworms, insect larvae, adult insects, and also some plant material.

Small burrowing animals are abundant in the forests of the northeastern states and have a pronounced influence on soil conditions. Hamilton (1940) indicated that the looseness of the surface layers in many of the forests of this region is due in part to the scores of burrows made by mice and shrews. The burrows frequently form a veritable labyrinth of interconnecting tunnels. The tunnels allow for ready penetration of water and air, and the nests, dung, stored food, and dead animals all add organic matter

and fertility to the soil. Observations of Eadie (1939) on the abundance of tunnels made by Brewer's mole (*Parascalpos breweri*) in a forested area in southeastern New Hampshire furnish evidence of the activity of these animals. Along a newly cut road bank there were 220 surface tunnels in a distance of 492.2 ft., or an average of one tunnel every 2.2 ft.

Quantitative information concerning the population of small mammals in forests of the Northeast has been obtained by Hamilton and Cook (1940). In Table 8 are presented average numbers per acre for several species.

TABLE 8. SPECIES DISTRIBUTION OF SMALL MAMMALS
IN FORESTS OF THE NORTHEAST

(After Hamilton and Cook, 1940)

Species	Number per Acre	Unit Weight, grams	Total Weight per Acre, grams
Deer mouse	35	20	700
Red-backed mouse	12	21	252
Flying squirrel	1	60	60
Short-tailed shrew	40	18	720
Long-tailed shrew	9	6	54
Hairy-tailed mole	3	54	162
Total	100		1948

Populations such as that indicated in Table 8, although not unusual, must be regarded as large. The total population of small mammals weighs 1948 grams, or 4.29 lb., per acre. Hamilton and Cook have pointed out that even on good range the weight of white-tailed deer averages less than two-thirds this amount (one 125-lb. deer per 50 acres). The relation of small mammals to the forest floor is shown diagrammatically in Figure 10.

Work of the pocket gopher and other rodents common in the western United States has been emphasized by Grinnell (1925, 1933) and Taylor (1935). Grinnell compared the role of burrowing rodents in western America to that of earthworms in more humid regions. Rodent tunnels are frequent in the longleaf pine (*Pinus palustris*) soils of the South which have been protected from fire, and in soils supporting eastern white pine (*Pinus strobus*) in the Northeast (Figure 11). In the well-drained sandy

soils of Florida the pocket gopher (*Geomys* sp.) accomplishes a considerable amount of soil cultivation.

Soil formations known as *krotovinas* have been studied rather

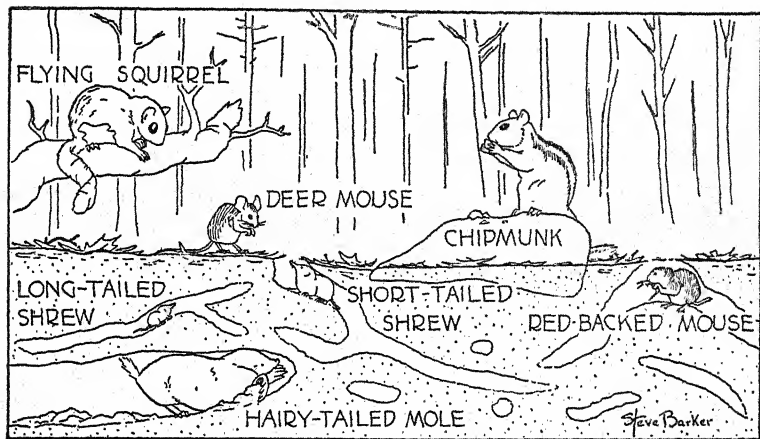


FIG. 10. The relation of small mammals to the forest floor. Courtesy, Hamilton and Cook (1940).

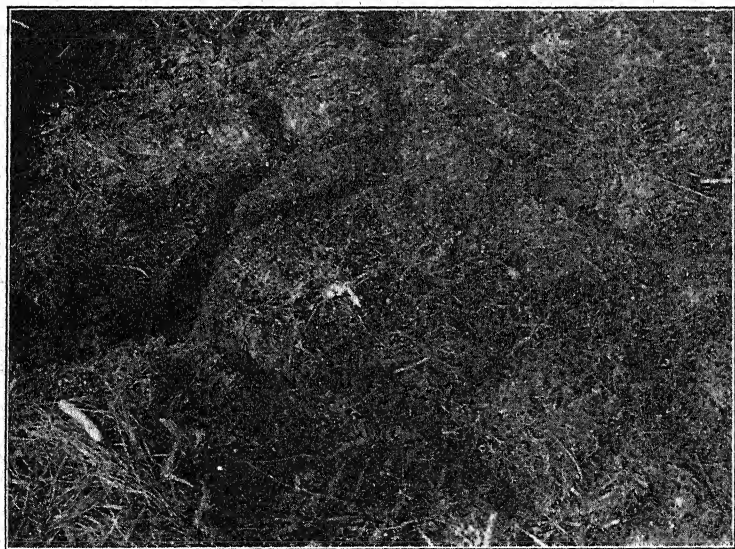


FIG. 11. Tunnels formed by small rodents in the A_1 horizon of a forest soil (Yale Forest, near Keene, New Hampshire). The layers of unincorporated organic matter have been removed.

extensively in the grassland regions of Russia (Joffe, 1936: 110). The term refers to the irregular formations resulting from the filling in of rodent tunnels. Soil material settles downward into passageways and nests, producing krotovinas, which appear in the profile as rounded spots or elongated streaks.

ARTHROPODA

Members of this phylum comprise a large proportion of the forest soil fauna. Some of the more important classes are considered below.

The class Crustacea is represented by "sow bugs" and "pill bugs" (order, Isopoda), which are quite prevalent in forest soils. These small animals (Figure 12, 9) are quite active, feeding on dead leaves and wood, and may be one of the agents responsible for the disintegration of freshly fallen leaves. Crayfish (order, Decapoda) appear to play a role in the poorly drained areas of the Southeast (Shaler, 1891; Heyward, 1937). Shaler maintained that the tunnels and galleries of these animals facilitate drainage of excess water and favor entrance of air into the soil.

The class Arachnida includes the mites (order, Acarina), false scorpions (order, Chelonethida), and spiders (order, Araneae). Members of the Chelonethida and Araneae are predaceous and do not play a direct part in the reduction of forest tree litter. Mites constitute the most important group of Arachnida by virtue of their great abundance and the fact that usually more than four-fifths of them consists of saprophagous species, the remainder being predaceous. The saprophagous mites feed on decaying leaves, wood, fungal hyphae, spores, and feces of other animals. Two species of mites and a false scorpion are shown in Figure 12, 1, 2, and 3. Jacot (1939) has observed that certain species of mites lay their eggs in partly decomposed needles of red spruce and balsam fir (*Abies balsamea*). The larvae which hatch proceed to eat the needles from the inside, until only the epidermal shell remains. They pass through three nymphal stages, finally emerging as adults. An adult mite and a nymph in a spruce needle are shown in Figure 13.

The abundance of mites in the humus layers of forest soils has been demonstrated by many investigators (Pillai, 1922; von Pfetten, 1925; Bornebusch, 1930, 1932; Jacot, 1936; and Eaton

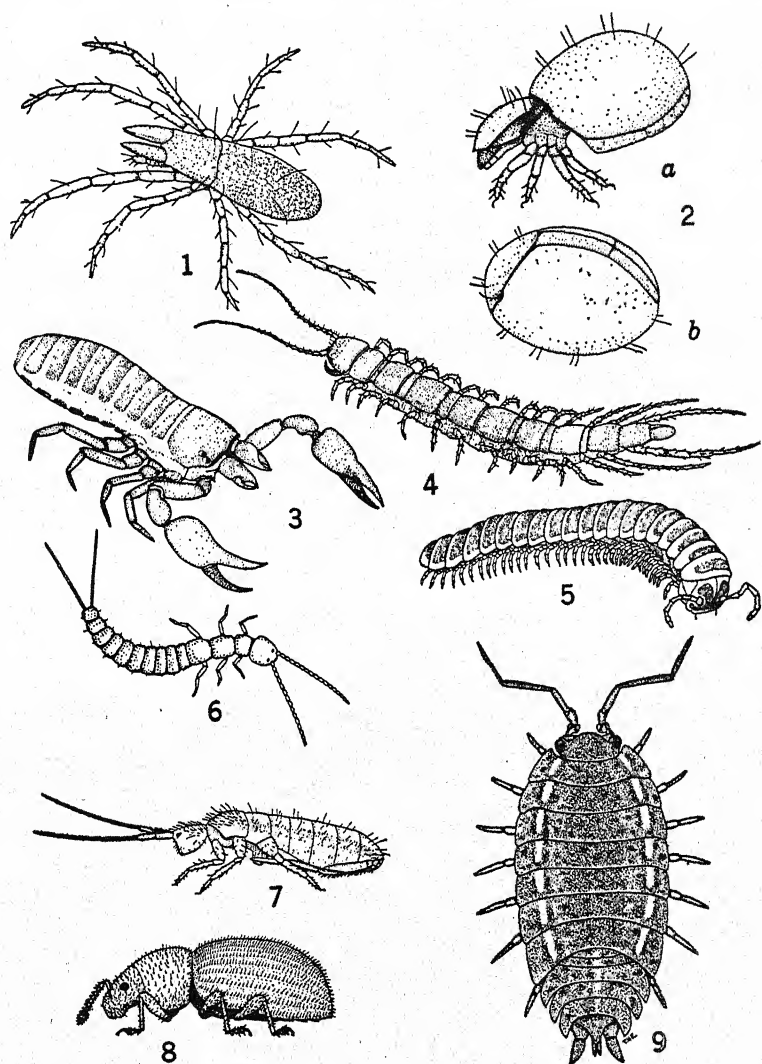


FIG. 12. Animals inhabiting forest soils. Courtesy, Eaton and Chandler (1942). 1. *Rhagidia*, a predatory mite, length 1 mm. 2. *Hoplodermas*, a saprophagous mite, length 0.5 mm.; (a) walking; (b) rolled up for defense. 3. *Neobisium*, a false scorpion, length 3 mm. 4. *Lithobius forficatus*, a common red centipede, length 35 mm. 5. *Fontaria coriacea*, a millipede, length 40 mm. 6. *Campodea staphylinus*, a spring-tail, length 4 mm. 7. *Tomocerus*, a spring-tail, length 4 mm. 8. *Barypeithes pellucidus*, a brown snout-beetle, length 3.6 mm. 9. *Trachelipus rathkei*, a sow bug, length 10 mm.

and Chandler, 1942). These animals are exceedingly numerous, the number per square foot commonly reaching several thousand. The only other arthropods that even approach the mites in abundance are the spring-tails, a group which will be considered

presently. Mites, as a group are very influential in promoting the breakdown of organic matter in forest soils and in regulating the course of decomposition.

The class Chilopoda includes the centipedes (Figure 12, 4), which are predatory animals feeding on many other members of the soil fauna. The class Diplopoda includes the millipedes (Figure 12, 5) which are chiefly saprophagous; in some forest soils they are very important members of the faunal population. Millipedes of the genus *Fontaria* are known to favor development of mull humus layer types in America (Romell, 1935a). In the Yale For-

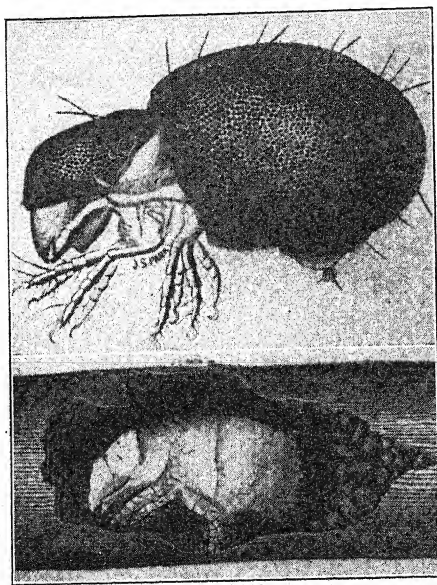


FIG. 13. The mite *Hoplophorella thoreauti*. Above, an adult much enlarged (natural size about 0.5 mm.); below, nymph in a spruce needle that has been broken open. From Jacot (1939); courtesy, *Journal of Forestry*.

est, near Keene, New Hampshire, millipedes are abundant in soils having excellent mull humus layers (Figure 14). In northern New England millipedes of the genus *Diploium* become active immediately after frost disappears from the soil in the spring but grow inactive during July and August. During the latter part of August they again become active and continue so until the soil freezes (Lyford, 1943). Feeding tests by Lyford demonstrated that leaves of certain tree species are definitely more palatable than others; in general, leaves of high calcium content are more palatable than those low in this element. Tree leaves showing the effects of feeding by millipedes are presented

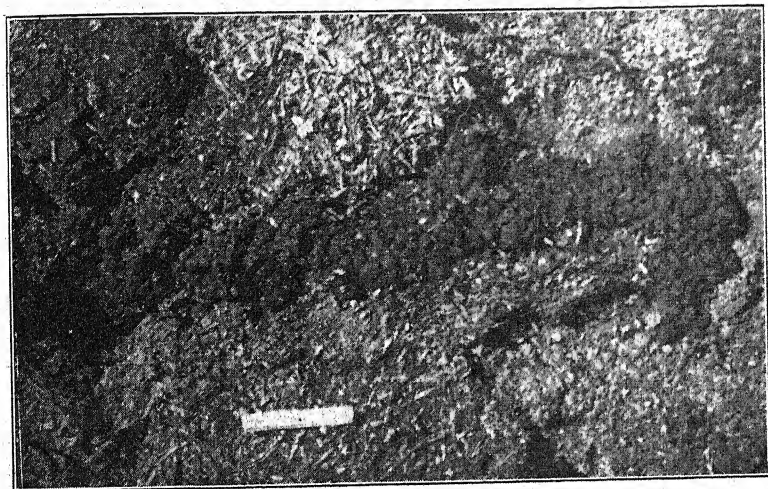


FIG. 14. Casts formed by millipedes in a soil supporting a pure stand of eastern white pine (Yale Forest, near Keene, New Hampshire). The *L* and *F* layers have been removed.

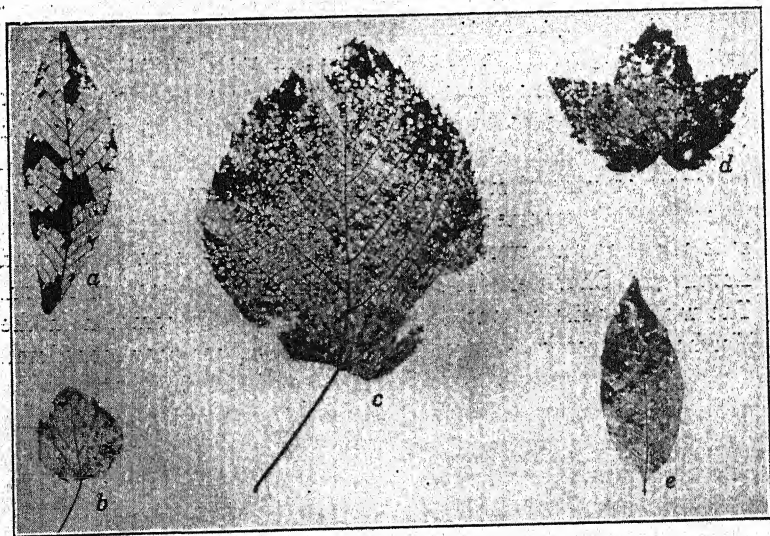


FIG. 15. Tree leaves showing effects of feeding by millipedes. *A*, Shagbark hickory; *B*, trembling aspen; *C*, basswood; *D*, red maple; and *E*, white ash. Photograph by Walter H. Lyford.

in Figure 15. Although millipedes assume considerable importance locally, it appears that as a rule they are not nearly as important in mull formation as are earthworms (Eaton and Chandler, 1942).

The class Hexapoda includes the true insects. Many representatives of this broad group occur in forest soils, but Collembola (spring-tails) occupy first place from the standpoint of number of individuals; with the exception of mites, they are usually the most abundant soil arthropods. The spring-tails are minute, wingless, saprophagous insects (Figure 12, 7). Another widely distributed saprophagous soil insect is the bristle-tail (Thysanura). It is small and wingless and bears two or three bristles posteriorly (Figure 12, 6). Adults and larvae of many beetles (Coleoptera) and larvae of various flies (Diptera) occur in forest soils; these organisms, being saprophagous, aid in the breakdown of organic debris and influence porosity by tunneling in the humus layer. The larvae of moths (Lepidoptera), although phytophagous and not regular inhabitants of the soil, contribute to the organic matter of the soil through their feces.

Ants (Hymenoptera) may play an important role in some soils because of their tunneling and transporting activities. It appears that their effect on forest soils is greater in tropical and warm-temperate regions than in temperate and cool-temperate regions. R. L. Donahue has observed the activity of ants in the rain-forests of Brazil and has noted that the genetic horizons in soil profiles are, in some cases, completely obscured as a result of vertical transport of mineral material by these organisms. Heyward and Tissot (1936) reported a much larger ant population in forest soils of the southeastern United States than Johnston (1936) and Eaton and Chandler (1942) found in the northeastern region. The mound-building ant (*Formica exsectoides*), common in the Northeast, must be regarded as harmful rather than beneficial. Trees are frequently killed for distances of 25 ft. or more around individual mounds. Damage is most frequent in plantations established on abandoned agricultural land, but injury may also be observed in natural stands.

MOLLUSCA

The members of this phylum which are of importance in soils belong to the class Gastropoda (slugs and snails). Bornebusch

(1930) found a number of these animals in Danish forest soils. He regarded *Arion subfuscus* as the chief species in the soils investigated. Blake (1931) recorded the occurrence of several species of snails in the soil of an American elm (*Ulmus americana*) and sugar maple forest in Illinois, and Strandine (1941) investigated the *Succinea ovalis* population in a hardwood forest on an old flood plain of the Desplains River. Most of the species found in soil consume organic debris, but some are carnivorous. It was pointed out by Shimek (1930) that land snails, particularly the larger forms, are distinctly forest inhabitants. He stated that the distribution of the larger land snails in the Mississippi Valley coincides rather closely with that of deciduous forests.

ANNELIDA

This phylum includes the segmented worms, of which the earthworms are the most important. Unfortunately, earthworms are commonly considered collectively, rather than as individual species. This is the practice in Europe, Heymons (1923) has pointed out, as well as in America. Heymons stated that frequently the earthworms are referred to collectively as *Regenwurmer*, in spite of the fact that there are over twenty species in Germany alone. Bornebusch (1930) has also recognized that different species vary considerably in importance.

Lumbricus terrestris is one of the most important species in Europe (Darwin, 1882; Müller, 1887; Heymons, 1923; Bornebusch, 1930). This large earthworm also occurs in America. Smith (1928) observed a tendency toward its increasing domination in Illinois and a corresponding decrease in the abundance of indigenous species; this same trend is reported in other parts of the world. Smith regarded *Lumbricus terrestris* as a European, not an indigenous, species. In America, *L. terrestris* is associated with good mull soils, a relationship which seems to exist in Europe (Müller, 1887). Bornebusch (1930) noted that *Allolobophora chlorotica* and *A. caliginosa* are also found in good mull soils. It appears that *L. terrestris* and other large species burrow deeply, whereas *Allolobophora* spp. tend to work near the surface. In the best mull soils the Lumbricidae make up about 80 per cent of the total weight of the soil fauna. In the surface mull in spruce stands Bornebusch commonly encountered such species as

L. rubellus, *L. castaneus*, and *Dendrobaena* spp. Species of *Dendrobaena* were most abundant in mor soils; in Norway spruce

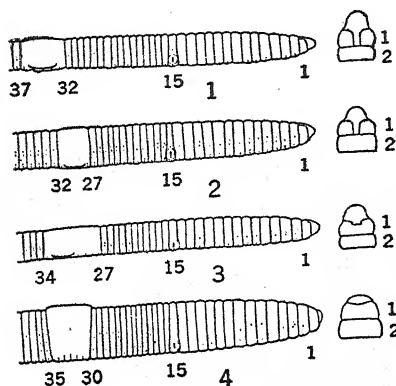


FIG. 16. Recognition characteristics for adults of four common earthworms in the northeastern United States. Courtesy, Eaton and Chandler (1942). (Lateral aspect, and dorsal view of anterior end. The small numbers indicate segments, counted from the mouth.) 1. *Lumbricus terrestris*, length 125-200 mm. 2. *Lumbricus rubellus*, length 100-150 mm. 3. *Allolobophora caliginosa*, length 100-175 mm. 4. *Octolasion lacteum*, length 60-100 mm.

mor, *D. octoedra* was usually encountered, but the number of individuals was low. In general, the number of earthworms in mor soils is much less than in mull soils. Eaton and Chandler (1942), working in New York state, found *Lumbricus terrestris* and *Allolobophora caliginosa* abundant in all forest soils having a coarse mull humus layer. *Lumbricus rubellus* and *Octolasion lacteum* occurred only occasionally. The characteristics by which these earthworms may be recognized are shown in Figure 16.

Enchytraeids, commonly called potworms or white worms, were reported by Waksman (1932) to be especially common in moist soils containing organic mat-

ter. They are small worms, mostly not more than a few millimeters long. Bornebusch (1930) found them about as numerous in mor as in mull, and Blake (1931) reported them extraordinarily abundant in the soil of an American elm and sugar maple forest in Illinois. Out of sixteen sites examined in New York state Eaton and Chandler (1942) found potworms in eight. These organisms, like the larger earthworms, devour organic debris and thereby assist in its decomposition (Jegen, 1920). Furthermore, they improve the physical properties of the soil. In view of the relatively small size of the enchytraeids, however, and the fact that their activity is concentrated in the uppermost soil layers, they are probably not as important as the earthworms.

Powers and Bollen (1935) calculated that in a Sites silty clay

loam soil in Oregon there were about 250,000 earthworms per acre. The land originally supported Douglas fir (*Pseudotsuga taxifolia*) but had been cut over. In mixed hardwood forest soils in Sweden, Lindquist (1938: 268) estimated an average of about 150 earthworms per square meter (approximately 605,000 per acre). Eaton and Chandler (1942) found that coarse mull soils in New York contained from approximately 522,000 to 1,010,000 earthworms per acre.

One of the first to recognize the importance of earthworms in the soil was White (1789: 216-217), who stated: "... worms seem to be the great promoters of vegetation, which would proceed but lamely without them, by boring, perforating, and loosening the soil, and rendering it pervious to rains and the fibres of plants, by drawing straws and stalks of leaves and twigs into it; and, most of all, by throwing up such infinite numbers of lumps of earth, called worm-casts, which, being their excrement, is a fine manure for grain and grass." Continuing, he expressed the view that "... the earth without worms would soon become cold, hard-bound, and void of fermentation; and consequently, steril ..."

Without doubt the most widely known work on earthworms is that of Charles Darwin. In 1840 his first report on the subject was published, being based on an address delivered before the Geological Society of London in 1837. He pointed out that cinders and marl applied to pasture fields years earlier had been buried by fine soil material brought up by earthworms. There followed a long period of observation and experiment, and in 1881 Darwin's classic work on the subject appeared.¹ He calculated that in England in the course of 50 years all the upper 10 in. of soil in a pasture had been brought up by earthworms. The general review of the Oligochaeta by Stephenson (1930) probably represents the best modern monograph on the subject.

Earthworms are beneficial, since they consume considerable quantities of organic debris and pass it, together with fine mineral soil particles, through their bodies. The resulting casts improve the physical nature of the soil. Earthworms also draw bits of organic matter into the mineral soil, and their burrowing activities promote good soil structure, good aeration, and good penetra-

¹ Cited from the American edition of 1882.

tion of water; their tunnels furnish avenues for development of plant roots. In 1890 Wollny published the results of his investigations of earthworms, carried on over a period of 7 years. He showed clearly that the presence of earthworms in soil results in a marked increase of pore space and air capacity, amounting to 27.5 and 22.3 per cent, respectively. Müller (1887) and Bornebusch (1930) concluded that the deep and friable mull of Danish deciduous forests is contingent on the activity of earthworms and associated species. Among others, Gast (1937) and Heyward (1937) have pointed out the value of these animals in American forest soils.

In addition to their physical influence it is evident that earthworms effect in the soil changes of a chemical nature. Hensen (1877, 1882) suggested the importance of earthworms from the standpoint of soil fertility, and Wollny (1890) obtained data which seem to show that the amount of soluble mineral matter in soil is increased by the activity of these animals. Hesselman (1937: 690) found that, when specimens of *Dendrobaena octaedra* are added to cultures of mor humus layer material, an increase in pH develops and, furthermore, that there is a marked increase in nitrogen transformation in the form of ammonia. Obst *et al.* (1930: 150) stated that the excrement of earthworms contains more soluble phosphoric acid and more ammonia and nitrate than the original soil. Lunt and Jacobson (1944) have recently shown that the chemical composition of earthworm casts from hardwood forests in Connecticut differs greatly from the composition of the adjacent soil material. The casts from hardwood forests contain more total nitrogen, carbon, available phosphorus, exchangeable and total calcium, and exchangeable magnesium and potassium. They also have higher cation-exchange capacity, base saturation, and pH. The excrement of earthworms and other animals forms a very favorable medium for the development of microscopic organisms (Francé, 1921: 86; Stoklasa, 1930).

The ecological requirements of earthworms are, in general, poorly understood. Various writers, for example, have discussed the influence of soil acidity, but there is a notable lack of agreement on the critical values. In 1899 Ramann *et al.* carried out simple experiments which led them to believe that earthworms can live in rather acid soils. They explained the relatively low numbers in soils having mor types of humus layers by the dryness

of the soil and by the pronounced changes in moisture content of the upper layers. Moore (1922) found earthworms in soil having a pH of 5.5, and Wherry (1924) observed the species *Helodrilus lönnbergi* in soil of pH 4.7. Waksman (1932) stated that earthworms are most abundant at pH 7.0-7.8 but occur in the range 5.8-8.3. The lack of agreement among these investigators on the limiting values of pH is not surprising in view of the fact that differences existed in species and in environmental conditions other than acidity. Although the exact acidity tolerances cannot be specified, it appears clear that, in general, highly acid soils support fewer earthworms than do less acid soils. It also seems clear that coarse-textured soils are relatively unfavorable, as also are soils which tend to be excessively dry (Diem, 1903: 153-154; Ramann, 1911a). Severe forest fires are probably unfavorable for these organisms (Jacquot, 1904: 309-310).

The depth to which earthworms descend in the soil is highly variable, depending upon the species, the soil, and the weather. During cold or dry weather they penetrate more deeply than when the soil is warm and moist. Müller (1887: 18) found earthworm galleries extending to a depth of 3-4 ft. In the temperate zone, earthworm activity shows two yearly maxima, one in the spring and another in the fall; during the winter and the hottest part of the summer activity is at a minimum.

In regard to food habits Gast (1937) cited Johnston as finding that leaves of the big-tooth aspen (*Populus grandidentata*), white ash, and basswood are readily eaten by *Lumbricus terrestris*. Leaves of sugar maple and red maple are eaten less readily, and northern red oak (*Quercus borealis*) leaves are not eaten.

TROCHELMINTHES

This phylum includes the rotifers or wheel animalcules; they are minute, mostly microscopic, animals. Waksman (1932) stated that they are commonly found in forest soils but comparatively little is known of their influence.

NEMATHELMINTHES

The members of this phylum are threadworms or roundworms. Nematodes, or true roundworms, are common inhabitants of forest soil; they occur in greatest abundance in the upper layers.

A favorable moisture supply is one of the most important conditions for their activity. Pillai (1922) found large numbers of these organisms in the Scotch pine soils of Bavaria, and von Pfetten (1925) found them in Norway spruce soils in the same region. According to Waksman (1932: 331), they are either parasitic on animals and plants or are saprophytic and free-living. The genera *Iota* (*Hoplolaimus*) and *Alaimus* are mentioned specifically as occurring in forest soils. Nematodes may influence soil conditions in the following ways: (1) consumption of soil protozoa, bacteria, and fungi; (2) distribution of bacteria and fungi through the soil; (3) transformation of soil organic matter; and (4) improvement of soil aeration.

PLATYHELMINTHES

The Turbellaria, or flatworms, are represented by various species in moist soils; they feed on other low forms of soil animal life. Trematoda, or flukes, also occur in soils.

PROTOZOA

Protozoa are the most elementary forms of animal life in the soil. They exist in both an active stage and in cyst condition. Numerically, they are the most abundant group of soil fauna. Fehér (1929b) reported 6250-9042 protozoa, encysted and active, per gram of moist soil from a Norway spruce forest. Waksman (1932: 675) stated that the number of protozoa in forest soils varies from 2500-10,000 per gram and that for the most part they are found encysted. Working with forest soils in the vicinity of Sopron, Fehér and Varga (1929) and Varga (1933) found two maxima in numbers of protozoa; the principal maximum came in late autumn (November and December) and a second, somewhat lower maximum, at the beginning of summer. Active protozoa were found at all times of the year; no differences could be noted in the protozoa of soils supporting hardwoods and those supporting conifers.

Most protozoa are aerobic and occur in the upper 4-6 in. of soil. The food of these unicellular animals consists largely of bacteria and decomposing organic matter. In general, the conditions which favor bacteria also seem to favor the development of protozoa (Francé, 1921; Waksman and Starkey, 1931: 70). Favorable soil moisture conditions are of outstanding importance.

Varga (1933) expressed the view that the significance of protozoa has in the past been underestimated.

GENERAL VIEW OF THE SOIL FAUNA

The more important representatives of the fauna of forest soils having been considered, certain general relationships between animal populations and soil conditions will now be discussed.

Forest soil humus layers are commonly separated into two broad groups, mull and mor. Mull represents a humus layer consisting of mixed organic and mineral matter, and mor refers to a humus layer of unincorporated organic material which is distinctly delimited from the mineral soil. Other characteristics of these two broad groups are presented in Chapter 6, but for present purposes the distinction just given is sufficient. In order to illustrate the differences in fauna between soils having mull and those having mor humus layers, the data in Table 9 have been compiled from the work of Eaton and Chandler (1942). The coarse mull represents a very fertile and productive soil having a pH of 6.75; moisture conditions are favorable. The matted mor has a pH of 4.30; as a result of a shallow, excessively drained soil, the site is subject to periods of extreme dryness during the summer.

TABLE 9. FAUNA OF TWO FOREST SOILS IN NEW YORK STATE (Values represent numbers per square foot to a depth of 1 in.)

Animals	Coarse Mull	Matted Mor
Earthworms	23*	0*
Potworms	80	27
Mites	865	2815
Spring-tails	265	895
Bristle-tails	1	5
False scorpions	3	35
Spiders	6	10
Centipedes	0	5
Millipedes	22	0
Symphylids	5	30
Isopods	55	0
Beetle larvae	14	49
Beetle pupae	5	4
Beetle adults	19	5
Fly larvae	15	17
Ants	1	63

* Represents entire depth of soil body, rather than 1 in.

The organisms listed in Table 9 are represented in the fauna of most forest soils of the temperate zone. Additional species of lesser importance also occur but were not included by Eaton and Chandler because of the sampling technique used. The principal difference in the populations indicated in Table 9 is the abundance of earthworms in mull and their complete absence in mor. Furthermore, the mite and spring-tail populations in mor are over three times those in mull. This general relationship has also been demonstrated by Francé (1921: 65), Bornebusch (1930, 1932), and Johnston (1936). Thus, mor humus layers generally contain the larger number of individuals, but mull humus layers usually have the greater weight of fauna because of the presence of earthworms and large arthropods. Bornebusch (1930, 1932) showed that the weight of animals in the poorest mor soil is only one-fifth that in the best mull soil.

Animal populations of forest soil vary considerably from one locality to another. This fact makes it difficult to arrive at average values for the various species. An attempt in this direction has been made, however, with the results given in Table 10. In arriving at these estimates, the data of Bornebusch (1930, 1932), Heyward and Tissot (1936), Johnston (1936), and Eaton and Chandler (1942) have been employed. Mites and spring-tails are present in all upland forest soils, but earthworms tend to occur chiefly in those having a mull humus layer. The es-

TABLE 10. ESTIMATES OF THE NUMBER OF CERTAIN ANIMALS IN MULL AND MOR HUMUS LAYERS OF FOREST SOILS (Values represent numbers per square foot to the depth indicated)

Humus Layer Type	Average Depth of Soil Sampled, inches	Acarina (Mites)	Collembola (Spring-tails)	Oligochaeta (Earthworms)	All Others*
Coarse mull	5.5	2240	745	12	310
Matted, granular, and greasy mor	3.4	4150	1335	1	130

* Except protozoa.

timates in Table 10 indicate that the mite and spring-tail population of mor humus layers may be approximately twice that found in mull, whereas there are about ten times as many earthworms

in mull as in mor. Animals representing groups other than the three mentioned are generally more abundant in mull than in mor. Furthermore, Eaton and Chandler (1942) found that the number of species occurring in mor is lower than in mull, particularly in very acid mor types developed under red spruce-hardwood and red spruce flat forest types of the Adirondack Mountains.

Data on the vertical distribution of fauna in a soil profile are presented in Table 11. These data are based on a well-developed podzol profile with a very acid, greasy mor humus layer which occurred beneath a stand of red spruce and balsam fir in the Adirondacks (Eaton and Chandler, 1942). The data show that

TABLE 11. DISTRIBUTION OF THE FAUNAL POPULATION IN THE SOLUM OF A PODZOL SOIL PROFILE (Values indicate numbers per square foot for the thickness of each layer)

Layer or Horizon	Depth, inches	Mites	Spring-tails	All Others*
<i>L + F</i>	0-2	15,000	1980	140
<i>H</i> (upper)	2-6	3,800	1420	20
<i>H</i> (lower)	6-10	2,400	500	20
<i>A</i> ₂	10-13	150	0	0
<i>B</i> ₁	13-16	120	30	0
<i>B</i> ₂	16-23	70	70	0

* Except protozoa.

the highest concentration of mites and spring-tails is in the *L* and *F* layers. In the mineral soil the number drops off very sharply. Johnston (1936) reported a considerably higher animal population in the *B* horizon of mull soils than in mor soils.

Although detailed information on the role played by forest soil fauna is very incomplete, certain general statements appear warranted. First, in temperate zone forests disintegration of the litter which falls to the ground each year is most prompt and incorporation of material in the mineral soil is most complete when earthworms or large millipedes are present in considerable numbers. On sites where these organisms are lacking, tree leaves that have lain on the ground several years may still retain sufficient structure to permit identification of their species. Second, the formation of coarse mull (see Figure 17) is directly associated with an abundant and active earthworm or large-

arthropod population. Third, saprophytic mites and spring-tails are beneficial because they break down partly decomposed leaves and other organic debris; these organisms do not, however, attack freshly fallen organic matter. Fourth, certain of the mites, spring-tails, and other arthropods feed upon fungal hyphae. In



FIG. 17. The forest floor of a hardwood stand having a coarse mull soil in which earthworms are abundant. Photograph taken in September just before leaf fall. Note that most of the leaves that fell during the previous year have been consumed, only the petioles and midribs remaining.

this way they perform an important function in forest soils, tending to prevent formation of excessively dense mats of hyphae. The occurrence of fibrous mor indicates, among other things, lack of a sufficiently active fungivorous fauna.

Whenever possible, the silviculturist should give consideration to the creation and maintenance of forest conditions favorable to an active soil fauna. Much additional research is needed on the ecology of soil organisms before it will be possible to state the desirable conditions with any degree of certainty. At present, however, it appears that the objective may best be attained by the development of full stands of healthy trees adapted to each site. Moderate crown density, frequent thinnings, an herbaceous ground cover, and elimination of repeated fires are all conditions

now believed to be favorable to the development of the most important forest soil animals.

FLORA OF FOREST SOILS

BACTERIA

Bacteria play an important role in forest soils. The following relations between microorganisms and higher plants have been pointed out by Waksman (1932: 763):

1. Plants and microorganisms influence each other by production of certain growth-promoting substances.
2. As a result of the activities of one group, substances are formed which are direct nutrients for the other group.
3. The two groups may actually live in symbiotic relation.
4. The two groups may compete with each other for certain nutrients in the soil.
5. One group may become injurious to the other by direct attack or by the action of toxic substances produced.

The number of bacteria in forest soils is highly variable; it may range from a few hundred thousand to many millions per gram of soil material. As a rule, the number of organisms in forest soil is considerably less than in agricultural soil (Waksman, 1916; Fehér, 1929*b*). As an indication of the relative numbers of microorganisms which have been found in forest soils, a few examples may be cited. Average numbers of organisms per gram of certain German forest soils, as reported by Ramann *et al.* (1899), are presented in Table 12. Cobb (1932) found 9,000,000

TABLE 12. NUMBER OF ORGANISMS IN GERMAN FOREST SOILS

(After Ramann *et al.*, 1899)

Kind of Soil	Bacteria	Fungi
	Number per gram	
Mull loam soils	4,860,000	2,339,000
Mull sand soils	2,500,000	3,404,000
Raw humus (mor) soil	247,000	3,320,000

bacteria per gram of deciduous forest soil and 6,660,600 per gram of eastern hemlock (*Tsuga canadensis*) forest soil. The data relate to counts made in April; samples consisted of material from

the organic layers. Powers and Bollen (1935) investigated the microorganisms in two Oregon soils supporting Douglas fir and reported the data which are presented in Table 13.

TABLE 13. NUMBER OF MICROORGANISMS IN OREGON FOREST SOILS

(After Powers and Bollen, 1935)

	Numbers per Gram, Water-Free Basis		
	<i>Bacteria</i>	<i>Actinomycetes</i>	<i>Molds</i>
<i>Aiken silt loam — field condition, not incubated</i>			
Litter	13,600	13,600	129,200
F	1,208,000	490,000	315,000
H	11,573,000	261,000	293,000
A, 0-6 in.	667,000	95,000	65,300
A, 6-15 in.	3,672,000	612,000	53,000
B, 15-36 in.	2,448,000	1,360,000	19,000
C	392,000	10,000	1,600
<i>Olympic silt loam — field condition, not incubated</i>			
Litter	66,780,000	882,000	10,100
F	47,538,000	11,628,000	85,500
H	34,542,000	9,234,000	171,000
H, transition	7,627,000	4,734,000	65,800
A ₁	1,980,000	642,000	37,500
A ₂	4,323,000	732,000	12,000
B	596,000	78,000	1,400
C	175,000	80,000	3,000

The largest number of bacteria is usually found in the surface soil (Waksman and Starkey, 1931: 46; Cobb, 1932; Fehér, 1933a: 60). Waksman (1916) investigated the vertical distribution of bacteria in a forest soil near New Brunswick, New Jersey, with the results indicated in Table 14. Powers and Bollen (1935)

TABLE 14. NUMBER OF BACTERIA IN A FOREST SOIL AT DIFFERENT DEPTHS

(After Waksman, 1916)

Depth, inches	Average Number of Bacteria per Gram of Dry Soil
1	2,088,000
4	1,172,000
8	482,000
12	311,000
20	169,000
30	104,000

found the greatest numbers of all classes of microorganisms in the humus layers and A horizon. In forest soils whose uppermost horizons are subjected to excessive drying out and intense solar radiation the greatest number of bacteria may be found at a depth of 5-15 cm. (Rippel, 1931: 257-258).

The number and kinds of bacteria present in a given soil are continually fluctuating. These fluctuations are of two types: (1) daily or even hourly changes, and (2) seasonal changes. It should be recognized that fluctuations in the population of soil organisms result in fluctuations of associated phenomena. Fehér (1929a; 1933a) investigated the seasonal changes of bacterial numbers in forest soils in central and northern Europe and found that the maximum occurred during the summer and the minimum in winter. A secondary maximum was reached in either the spring or the autumn. The maximum number of nitrifying bacteria occurred in the spring. Denitrifying bacteria were most abundant in the spring in northern Europe, but in central Europe the highest numbers were encountered in the autumn and winter. Nitrogen-fixing organisms attained their maximum numbers in the fall. Cellulose-decomposing bacteria in soils under conifers were found to attain their maximum in the spring, with a secondary maximum in the late summer; under hardwoods the maximum came in the summer, with a secondary maximum in the spring. The fluctuations in the number of anaerobes were much less than in the number of aerobes; both were most abundant in the summer, least abundant in winter. Cobb (1932), working with forest soils in New York, found that the maximum number of organisms occurred in April. Lochhead (1924) reported that the numbers of bacteria and fungi in soil remained relatively constant during the winter; an increase in numbers was first noted when the soil began to thaw in the spring. A number of factors influence the seasonal fluctuations just noted, but soil temperature and soil moisture appear to be outstanding. Frequently moisture is deficient during the time when temperature conditions are most favorable; when moisture is abundant, the temperature may be too low.

The importance of favorable temperature relations for microorganisms has been emphasized by Süchting (1929), Sachsse (1932: 203), Fehér (1933a), and Lipman (1936). Sachsse stated that the most favorable temperature for bacteria active in the

decomposition of leaf material is about 35°C. Under temperate zone forest stands the prevailing soil temperatures are considerably below the optimum. Soil moisture in relation to microorganisms has been considered by Ramann *et al.* (1899), Süchting (1929), and Fehér (1933a). Bühler (1918: 392) and Dubos (1928) regarded the optimum soil moisture content as between 60 and 80 per cent of saturation, and Waksman (1932: 546) stated that optimum conditions prevail when about one-half of the pore space is filled with water. The non-spore-forming bacteria seemingly require higher moisture content for their best development than do the spore-forming bacteria (Fehér, 1933a).

The importance of soil acidity in relation to the development of bacteria has been emphasized by Ramann *et al.* (1899), Süchting (1929), and Waksman and Starkey (1931: 59). In general, high acidities are regarded as unfavorable. Fehér (1929b, 1933a) noted that frequently the fluctuations in pH of soils parallel fluctuations in the numbers of bacteria. However, he suggested that the variations in acidity may be a result, rather than a cause, of variations in the activity of the microorganisms.

It appears that the nutritional requirements of most microorganisms are similar to those of the higher plants. Good aeration, a source of mineral nutrient elements, a supply of easily decomposable organic matter, and good physical conditions of the soil (especially a high air capacity) all favor the development of most bacteria. Calcium appears to be particularly influential. Räuber (1914) found ten to twenty times as many bacteria in calcareous soils as in noncalcareous sandy soils. In some forest types burning may favor increased bacterial activity. Soil from burned-over longleaf pine land in the South is reported to contain greater numbers of bacteria than does soil from unburned areas (Greene, 1935). Good aeration is particularly important for the activity of nitrifying and nitrogen-fixing bacteria (Voss, 1929). For ammonification, aerobic conditions are desirable but not necessary; nitrification is a more sensitive process, and good aeration is essential (Schairer, 1937). Düggeli (1937) investigated the influence of trampling on forest soils near Zürich. Aerobic nitrifying bacteria and members of the *Azotobacter* group are less abundant in the compact soil than in adjacent undisturbed soil; denitrifying bacteria are more abundant.

It has been found that microorganisms occur in greater abun-

dance near plant roots than at a distance from them (Starkey, 1929, 1931, 1938). The bacterial population, which is more affected by roots than are the filamentous fungi and actinomyces, is highest on the root surfaces. This influence of plant roots has also been noted by Gräf (1930) and by Thom and Humfeld (1932); these authors suggested that the reaction in the vicinity of roots is more favorable than at some distance away.

Forest composition and density are factors which exert a pronounced influence on the microorganic population of the soil. Fehér (1929b), working in Hungary, found the highest numbers of bacteria in soils from mixed hardwood-conifer stands, lower numbers under hardwood stands, and lowest numbers under conifers. The researches of Svinhufvud (1936, 1937b) in Finland are particularly interesting. He reported that the quantitative occurrence of soil bacteria (aerobes and anaerobes) and fungi showed great differences in different forest (site) types. A rather direct relationship was found between the number of microbes and the excellence of site. The indicator types, together with the total number of bacteria, per gram of soil, found in September, follow:

Oxalis-Maianthemum	13,501,000
Oxalis-Myrtillus	7,926,000
Myrtillus	5,337,000
Vaccinium	4,871,000
Cladonia	3,095,000

Marked differences between the types in the various physiological groups of organisms were also noted.

Vandecaveye and Katznelson (1940) reported that the bacterial population is usually larger and the fungal population usually smaller in sierozem and grassland soils than in forest soils, particularly podzols.

AUTOTROPHIC AND FACULTATIVE AUTOTROPHIC BACTERIA

These organisms derive their carbon primarily from the carbon dioxide of the atmosphere and their energy from the oxidation of inorganic materials or simple compounds of carbon.

Bacteria Using Nitrogen Compounds as an Energy Source. Two groups of organisms will be considered here: (1) bacteria which oxidize ammonia to nitrite (*Nitrosomonas*, *Nitrosocystis*,

and *Nitrosospiras* groups), and (2) bacteria which oxidize nitrite to nitrate (*Nitrobacter* group). Organisms capable of oxidizing ammonia directly to nitrate are not known. Some 40 years of research has confirmed the view that nitrifying organisms are autotrophic and that their energy is based exclusively on the oxidation of ammoniacal nitrogen and nitrous nitrogen (Winogradsky and Winogradsky, 1933: 428). It is unproved that heterotrophic microorganisms are capable of nitrification.

Physiologically, from the point of view of their energy of action, the groups of nitrite organisms are arranged in descending order as follows: *Nitrosomonas*, *Nitrosocystis*, and *Nitrosospiras*. The first group includes the monad forms, and the second the "zoöglea" or cyst forms. The nitrite bacteria have an extremely variable distribution in forest soils. Evidently differences in acidity, the nitrogen relations in the media, and other factors account for the occurrence of these organisms. The nitrosomonades are the dominant or the exclusive representatives in fertile agricultural soils. *Nitrosocystis* species seem to be characteristic of nitrifying forest soils; nitrosomonades, if occurring at all, are rare. The *Nitrosospiras* form thus far has been found only in poor, uncultivated soils (Winogradsky and Winogradsky, 1933). A *Nitrosocystis* was found by Romell (1932) in crumb mull from a locality in New Hampshire and in root duff (matted mor) in New York. He suggested that the *Nitrosocystis* type is characteristic of nitrifying soils having either mull or mor types of humus layers. Romell (1927-28) earlier identified a similar nitrite organism in Swedish forest soils.

Organisms belonging to the *Nitrobacter* group oxidize nitrites to nitrates. The tolerance of these organisms in respect to acidity appears rather great, but the optimum is between pH 7.2 and 7.4 (Winogradsky and Winogradsky, 1933). The form *Bactoderma alba* is also a nitrate bacteria and has an optimum pH somewhat above that of *Nitrobacter*.

Fehér (1929b) stated that the number of nitrifying bacteria in forest soil is generally small and commonly far exceeded by denitrifying forms. High acidity was suggested as one of the reasons for the relatively low number of nitrifying bacteria in forest soils. It appears that the poorer is the soil type, the greater is the activity of the denitrifying forms (Svinhufvud, 1937a). In the opinion of Waksman (1932) all soils, if not too

acid, contain nitrifying bacteria. He placed the limiting acidity at pH 3.7-4.0, and the optimum at pH 6.8-7.3. The activity of the different groups of nitrite bacteria is not influenced in the same way by the acidity of the medium (Winogradsky and Winogradsky, 1933). For all forms investigated by these authors the optimum was between pH 6.0 and 9.2. In the majority, activity was nil below pH 6.0. Ammonification proceeds at rather low pH values, for example, 3.2-3.6 (Schairer, 1937).

Other ecological factors besides acidity also influence these bacteria. Gaarder and Hagem (1921) regarded soil moisture, aeration, temperature, and content of bases as highly important in determining nitrification. The most favorable moisture content is believed to be not less than 10-20 per cent nor more than 80 per cent of the water-holding capacity of the soil. Fehér (1933a) stated that with low air capacity of the soil the number of nitrifying forms decreases, and the denitrifying forms become dominant. The optimum temperature for most bacteria and fungi has been found to lie between 25° and 30°C. (Fehér, 1933b). There is reason to believe that the composition of the organic debris influences nitrification. Humus layers which have a high percentage of benzol-alcohol extractable matter usually show poor nitrification capacity (Němec, 1929). Nitrification proceeds most rapidly after almost complete mineralization of the organic matter. Olsen (1932) found that nitrification in beech and oak leaves does not begin until decomposition reaches a point where the nitrogen content is about 2.2 per cent (dry-matter basis). When the carbon-nitrogen ratio is greater than about ten, the nitrogen compounds liberated by decomposition are continually taken up by the microorganisms decomposing the carbon compounds. Gradually the carbon content decreases as a result of loss of carbon dioxide. When the carbon-nitrogen ratio becomes about ten or less, only part of the nitrogen liberated in available form during decomposition is used again by the microorganisms; the remainder accumulates as ammonia and nitrates.

Wiedemann (1924: 17) regarded nitrifying organisms as rather sensitive to high temperatures. Temperatures of 45°C. and above are lethal to nitrite-forming organisms, and nitrate-forming organisms are unable to withstand temperatures above 55°C.

From time to time it has been suggested that in some regions

heat and light may play important roles in nonbiological formation of nitrates. However, Waksman and Madhok (1937) investigated this question and concluded that "... the biological oxidation of ammonia to nitrite and nitrite to nitrate must still be considered as the all-important process in the formation of nitrate in soil."

Bacteria Using Sulfur and Sulfur Compounds as Energy Sources. Native sulfofying bacteria appear to be widely distributed; they were found in all mineral soil and peat samples examined by Wilson and Higbee (1932) in New York. The sulfur bacteria do not form a uniform group, as do the nitrifying bacteria (Waksman, 1932). They may oxidize hydrogen sulfide and other sulfides, elementary sulfur, or thiosulfate, and they may act in either an acid or an alkaline medium. The organism *Thiobacillus denitrificans*, which is widely distributed in soils, can oxidize various sulfur compounds and under anaerobic conditions may reduce nitrates to gaseous nitrogen. *Thiobacillus thiooxidans* does best in acid media, using free atmospheric oxygen in the oxidation of elementary sulfur to sulfuric acid. As a result, the hydrogen-ion concentration of the substratum may become very high.

Bacteria Using Iron and Manganese Compounds as Energy Sources. Bacteria capable of oxidizing ferrous iron to ferric iron, using the energy obtained for the assimilation of carbon, are regarded as the "true" iron bacteria (Starkey and Halvorson, 1927; Waksman, 1932: 89). They are separated from bacteria which may absorb or accumulate iron or which can affect precipitation of iron because of changes in the reaction of the medium. Iron is precipitated from solution by the organisms which live by oxidizing ferrous iron to ferric iron (Waksman and Starkey, 1931: 176). Oxidation of FeCO_3 to $\text{Fe}_2(\text{OH})_6$ is accomplished by various forms.

The occurrence of iron-precipitating organisms is universal; they have had an important part in the formation of bog iron ore. Harder (1919) stated:

In general it seems reasonable to believe that certain iron-depositing organisms, such as *Spirophyllum*, require ferrous bicarbonate in solution and cannot live without it; that others, such as *Leptothrix*, can live without any iron compounds, but if they are present can use either ferrous bicarbonate or soluble organic iron salts; and that still others,

such as various lower bacteria . . . will use the organic radical of certain soluble organic iron salts when present but cannot utilize any inorganic iron salts.

The deposition of iron by bacteria has also been discussed by Mudge (1927). In addition to iron compounds some bacteria oxidize manganese salts as well and precipitate manganese hydrates in their cells (Waksman, 1932).

The oxidation of ferrous compounds to ferric compounds by iron bacteria is analogous to the oxidation of manganous compounds to manganic compounds by manganese organisms.

HETEROTROPHIC BACTERIA

The heterotrophic organisms obtain their carbon and energy from various organic compounds.

Nitrogen-Fixing Bacteria. The nitrogen-fixing bacteria derive their nitrogen in gaseous form, from the atmosphere. It appears that none of them are obligate, since they can also obtain nitrogen from organic and inorganic nitrogenous compounds (Waksman, 1932: 102). Olsen (1932) has pointed out that the free-living nitrogen-fixing organisms can use ammonia or nitrates as a source of nitrogen. In fact, they seem to prefer these sources to atmospheric nitrogen. If ammonia or nitrates are present, little if any atmospheric nitrogen will be fixed.

Two groups of these bacteria may be recognized: (1) nonsymbiotic and (2) symbiotic. Of the nonsymbiotic forms the anaerobic types will be considered first. The great microbiologist Winogradsky (1893, 1895, 1902) isolated the first nonsymbiotic, nitrogen-fixing organism, *Clostridium pasteurianum*. In his 1895 report announcing isolation of the organism he stated that it had been named for Pasteur, creator of the science of microbiology. This anaerobic organism belongs to the butyric acid group of bacteria. The *Clostridium* forms are spore-forming and rod-shaped. Other organisms in this group capable of nitrogen fixation are species of the genus *Granulobacter*. It has been demonstrated that *Clostridium* has a very wide occurrence. Svinhufvud (1937a) reported it as occurring in all the Finnish soils he examined.

According to Waksman (1932: 672), the number of anaerobic nitrogen-fixing organisms in forest soils may vary from 100 to

10,000 per gram of soil; this same range was reported by Bokor (1927) for Swedish soils. It appears that the anaerobic forms are more common in forest soils than the aerobic types because of the generally high humus and moisture contents (Fehér, 1933a).

The optimum reaction for *Clostridium pasteurianum* was indicated by Waksman (1932: 107) as pH 6.9–7.3, but it may be found in soils having a pH as low as 5.0. Olsen (1932) reported that no fixation of nitrogen occurred below pH 5.0, and the amount fixed at pH 5.0–6.0 was negligible. The best results were obtained from pH 6.0 to 7.7. Consequently, he regarded conditions for nitrogen fixation as more favorable in forests with neutral or alkaline soil than in those having very acid soil.

The carbon-nitrogen ratio of forest tree leaves appears to influence the possibility of nitrogen fixation by free-living organisms. From this standpoint material having a high carbon-nitrogen ratio appears more favorable than does material having a lower ratio. Olsen (1932) was able to demonstrate nitrogen fixation in oak and beech leaves but not in leaves of hornbeam. Hornbeam leaves contain 1.75 per cent nitrogen (dry-matter basis) and decompose quickly, with the formation of ammonia. Nitrogen fixation ceases when nitrification begins.

The first of the aerobic types of the nonsymbiotic, nitrogen-fixing bacteria *Azotobacter* was isolated by Beijerinck (1901). *Azotobacter* organisms are large, coccus-like, and nonmotile. In addition to members of the *Azotobacter* group representatives of the *Radiobacter* group are also capable of fixing atmospheric nitrogen. The life history of *Azotobacter* has been considered in detail by Löhnis and Smith (1923).

Waksman (1932: 672) stated that aerobic nitrogen-fixing organisms are scarce in forest soils. This statement is supported by the fact that Weis and Bornebusch (1914) found *Azotobacter* in only 2 out of 64 Danish soils. In 106 samples of forest soil from northern Idaho *Azotobacter* were found in 24.5 per cent (Gibbs and Batchelor, 1927). Bokor (1927), working in Sweden, concluded that aerobic nitrogen-fixing bacteria were absent or of only limited occurrence in beech, alder, and pine forest soil. On the other hand, Svinhufvud (1937a) found *Azotobacter* in all the Finnish forest soils examined.

It appears that high acidity is one of the principal reasons for

the relatively low number of aerobic nitrogen-fixing organisms in forest soils. The investigations of Fehér (1933a) indicated that the *Azotobacter* population in Hungarian forest soils usually reached a maximum in the late summer or autumn, when the pH value of the soil was high. Gainey (1918) believed that pH 6.0 represented the lowest value which could be tolerated by *Azotobacter*, but Fred and Davenport (1918) placed the limit at about pH 6.5. In recent investigations Gainey and Fowler (1945) concluded that the maximum hydrogen-ion concentration compatible with growth of *Azotobacter* corresponded to pH 5.5-6.0. Winoogradsky and Ziemiecka (1928) stated that *Azotobacter* were generally absent in soils having a pH of less than 6.0 and in those lacking lime. The importance of lime is also indicated by the experience of Weis and Bornebusch (1914), who found a high content of calcium carbonate in soils containing the organisms. *Azotobacter* (*A. chroococcum* and *A. beijerinckii*) were found in Minnesota peat soils only when the pH was 5.9 or more (Skinner and Nygard, 1930).

Conditions other than acidity influence the occurrence and activity of *Azotobacter*. The supply of mineral nutrients, particularly potassium and phosphorus, and the available energy material are also important (Waksman and Starkey, 1931: 107). Mattern (1928) in Prussia found *Azotobacter* present in beech mor, but nitrogen fixation was low. She attributed the depression of activity to the presence of toxic or inhibiting substances. The relatively low soil temperatures in many forest stands may also be unfavorable.

The second major group of nitrogen-fixing organisms consists of the symbiotic forms. In 1888 Beijerinck isolated *Bacterium radiculicola* (*Pseudomonas radiculicola*), which he recognized as the agent causing nodules on the roots of leguminous plants. The organism is aerobic, and there are many different strains. According to Löhnis (1930), *Bacterium radiculicola* cannot assimilate nitrogen in the absence of the host plants.

Many species of the Leguminosae, and some plants of other families, are known to have nodules on their roots. Lists of these species have been prepared by Fred *et al.* (1932) and Bushnell and Sarles (1937). Occurrence of nitrogen-fixing bacteria in the nodules of nonleguminous plants has been reported occasionally. For example, Johnson (1917) stated that nodules from the roots

of red alder (*Alnus oregana*) contain these organisms. He maintained that alder exerts a very favorable influence on soil fertility in the Siuslaw National Forest in Oregon. In 1931 Phillips reported finding *Pseudomonas radicola* in the root nodules of four species of *Podocarpus* in South Africa. Among other non-legumes having root nodules *Ceanothus*, *Elaeagnus*, and *Myrica* may be mentioned.

In spite of the claims of Johnson, Phillips, and others evidence of the occurrence of the rhizobia of the Leguminosae in root nodules of plants of other families may be regarded as inconclusive (Burrill and Hansen, 1917; Waksman, 1932: 134; Fred *et al.*, 1932). It is probable that in certain forest types nitrogen-fixing organisms associated with the roots of wild legumes play an important role in the nitrogen economy of the soil. The importance of symbiotic nitrogen-fixing organisms is greatest in regions where native legumes are of common occurrence, for example, in the southeastern United States.

From the standpoint of soil reaction *Bacterium radicola* is not exacting. On the acid side the critical values appear to lie in the region of pH 3.2-5.0 and on the alkaline side near pH 9.0-10.0; optimum conditions occur within the range of pH 5.5-7.0.

Aerobic Bacteria Which Require Combined Nitrogen. This group includes a large number of organisms which are important in the decomposition of organic substances in the soil. Materials such as proteins, protein derivatives, celluloses, pentosans, and other carbohydrates are decomposed by bacteria in this group. The three most common spore-forming bacteria in soils are *Bacillus mycoides*, *B. cereus*, and *B. megatherium*; they are of universal occurrence (Waksman, 1932: 144). Waksman stated that the heterotrophic non-spore-forming bacteria are the predominant group of soil organisms. They appear to be important in the mineralization of the soil humus. Aerobic and anaerobic cellulose-decomposing bacteria are found in forest soils with a pH as low as 4.3.

Stapp and Bortels (1934) reported finding cellulose-decomposing organisms in all forest soils examined. The specific cellulose-decomposing organisms of the genus *Cytophaga* were found principally in the litter of hardwoods and occasionally in spruce litter, but never in pine litter. In explanation of this occurrence the authors suggest that the litter of spruce and pine

tends to be poor in bases and to have a low pH value. This may in part account for the frequently slow decomposition of spruce and pine litter. *Cytophaga crocea*, which occurred in base-rich hardwood leaves, was especially active in decomposing cellulose.

Anaerobic Bacteria Which Require Combined Nitrogen. Anaerobic bacteria are widely distributed and occur in great abundance in soils. Waksman (1932: 672) stated that the numbers vary from 500,000 to 5,000,000 per gram of forest soil. He also reported (p. 184) that the decomposition of cellulose under anaerobic conditions is carried on entirely by bacteria. It appears that the number of anaerobic pectin- and cellulose-decomposing bacteria stands in inverse relation to the absolute air capacity of the soil (Fehér, 1929*b*; 1933*a*).

Under anaerobic conditions many bacteria are capable of using inorganic salts rich in oxygen as a source of that element. Reduction processes such as the following may occur:

Nitrates $\xrightarrow{\text{(Reduction)}}$, nitrites, ammonia, nitrous oxide.

Sulfates $\xrightarrow{\text{(Reduction)}}$, sulfites, sulfides.

Phosphates $\xrightarrow{\text{(Reduction)}}$, phosphites, hypophosphites.

ALGAE

Algae are present in soils both as spores and in the vegetative condition. They are of universal distribution at the surface of soils, providing that moisture and light are available (Waksman, 1932: 201). Fehér (1936) reported a seasonal periodicity in the number of soil algae; abundant soil moisture favors an increase in the population. In coniferous forests in Hungary the species composition is commonly richer than that in soils supporting hardwood forest.

Gistl (1934) found algae to vary considerably with different soil conditions and suggested that some species are so sensitive they may be used as indicators of soil conditions. Stokes (1940*b*) reported that the optimum range of moisture conditions for algae is essentially the same as that for bacteria and *Actinomyces*, namely, 40–60 per cent of the moisture-holding capacity of the soil. It was found that the addition of organic matter to the soil inhibited the growth of algae during the period of active decomposition of the material by bacteria and *Actinomyces*;

Stokes suggested competition for essential elements as the cause.

At the soil surface the algae, since they bear chlorophyll, are capable of photosynthesis, but in the lower depths they are heterotrophic. The role played by algae is not clear, but it is likely that it consists in part of building up organic matter in young soils. Bühler (1918: 393) regarded algae as important in sandy soils, since they tend to bind the material. In the south-central United States several species of algae constitute an initial stage of plant succession on badly eroded lands. Soil erosion on plots supporting an algal cover is much less than on plots of bare soil (Booth, 1941). A few of the blue-green algae are reported to fix atmospheric nitrogen (Waksman, 1938; Stokes, 1940a). It appears that in general, the role of algae in soils is not very important.

FUNGI

Fungi are heterotrophic organisms which may occur in the soil both as mycelia and spores. McLennan (1928), working at Rothamsted, believed that the normal fungal constituents of soil are present extensively in the mycelial condition rather than as spores. Since soil fungi are dependent on the decomposition of organic compounds for their energy, they are of considerable importance in the alteration of these materials.

Waksman *et al.* (1928) and Waksman (1932: 670) regarded fungi in forest soils as more abundant, especially when the actual amount of microbial cell substance is computed, than bacteria. Fungi are especially abundant in soils rich in humus and having an acid reaction. In his studies of the microbiology of European forest soils Fehér (1933a: 160) found that the proportion of fungi to bacteria increased as he went north. In northern latitudes the fungi are particularly important because of the relatively low numbers of cellulose-decomposing bacteria.

The organic layers in an eastern hemlock stand in New York were found by Cobb (1932) to contain about twice as many fungi as the organic layers from a deciduous forest. On the average there were 3,948,000 fungi per gram of organic matter in the hemlock forest, and 1,716,490 per gram in the deciduous forest.

One of the chief functions of fungi is the decomposition of cellulose and related compounds (Jensen, 1931). The decomposi-

tion of woody residues under aerobic conditions is in large measure accomplished by fungi; they are particularly active in the early stages of decomposition of the organic debris of forest soils. Stüchting (1929) stated that fungi have the capacity of energetically decomposing organic nitrogen compounds to ammonia. In addition to decomposing organic material, fungi assimilate soluble inorganic nitrogen compounds and mineral elements. Fungus mycelium may contain from 3.5 to 8.0 per cent nitrogen. The work by Allison *et al.* (1934) indicates that fungi are incapable of assimilating atmospheric nitrogen.

Certain fungi, especially the Hymenomycetes, may form an association with plant roots. This association results in the formation of an organ called a mycorrhiza (fungus root). At least two well-defined types of mycorrhizae may be recognized, namely, (1) *ectotrophic*, in which the fungal elements occur in the intercellular spaces and also form a mantle which completely covers the root, and (2) *endotrophic*, in which the mycelium of the fungus is characteristically intracellular. In the endotrophic type no fungal mantle is present.

The ectotrophic type is found on all members of the Abietineae, Fagaceae, Betulaceae, and Salicaceae, as well as on members of individual genera of certain other families. Endotrophic mycorrhizae are even more common, being found in most terrestrial plants which have been investigated. Ectotrophic mycorrhizae, which are of primary interest to foresters, may usually be recognized in the field by the mantle of interlaced fungal hyphae covering the short roots. Branching is dichotomous in the pines but apparently racemose in other trees. Mycorrhizal and nonmycorrhizal roots of pine are illustrated in Figure 18.

Mycorrhizae appear to exert a favorable influence on plant development (Melin, 1927; Hesselman, 1927; Romell, 1935b; Hatch, 1937; McComb, 1943; Routien and Dawson, 1943). Melin found mycorrhizae best developed on the most vigorous pine seedlings in Sweden; he also found them better developed in the *F* than in the *H* layers. It was concluded that mycorrhizae benefit plants directly by assisting in nutrient absorption and indirectly by protecting the roots from harmful fungal mycelia. A lively transformation of nitrogen in mor humus types was reported by both Melin and Hesselman to favor the development

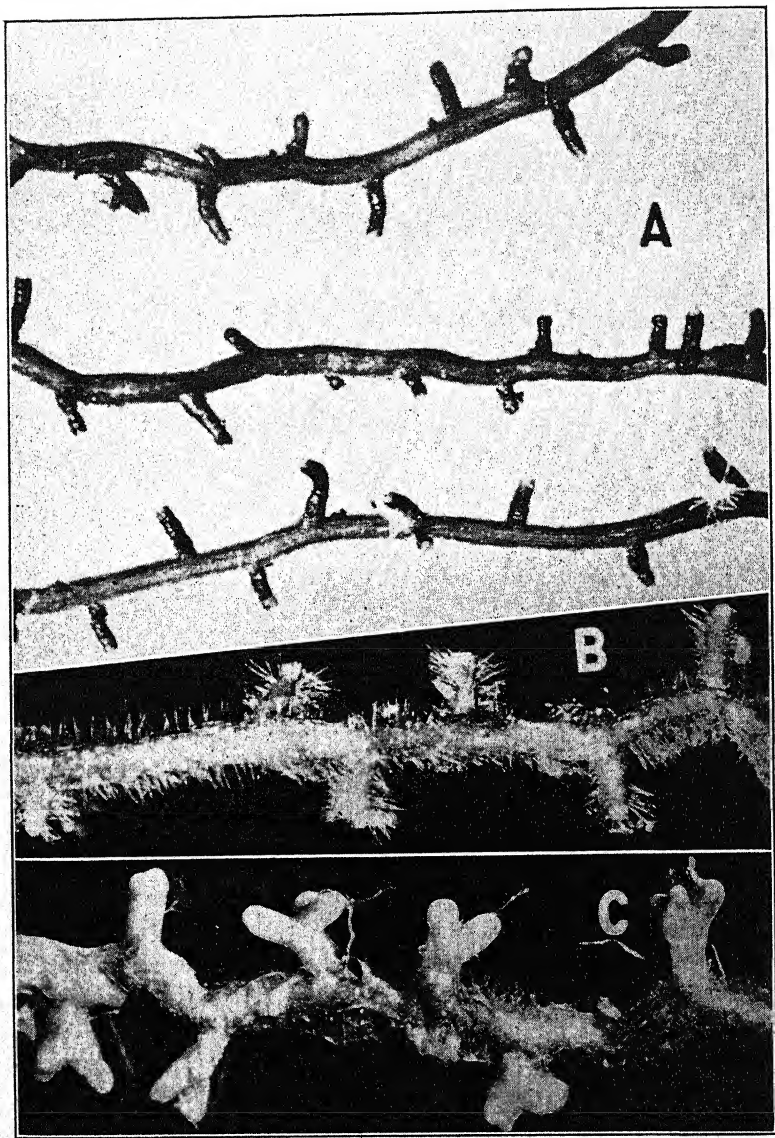


FIG. 18. Three types of short roots in pine. A, nonmycorrhizal; B, uninfected, but with root hairs (in pure culture); C, mycorrhizal. Courtesy, Hatch and Doak (1933).

of mycorrhizae and vigorous tree seedlings. Romell suggested that the beneficial effects of mycorrhizae arise, not as a result of attack on ordinarily unavailable sources of nitrogen, but "... by serving as additional roots helping in the general keen competition for available nitrogen food." Hatch considered the primary effect of mycorrhizae to be an increase of the absorbing surface of the plant and believed that they were particularly important in infertile soils.

MacDougal and Dufrenoy (1941) found that isolated segments of mycorrhizal pine roots were capable of independent growth for a period of 28 months. These authors reported that carbohydrates, adequate for growth for over 2 years, were synthesized by the mycorrhizal fungi. It was assumed that the mycorrhizae employed in the investigation were capable of synthesizing vitamin B₁, since synthesis of this substance "... has been found so widely distributed among fungi. ..." Recent researches by Routien and Dawson (1943) have demonstrated that development of mycorrhizae on shortleaf pine (*Pinus echinata*) seedlings increased the average rate of aerobic carbon dioxide production of each short root by nearly two to four times the normal amount. These authors suggested that the primary reason for the increased capacity of mycorrhizal roots to absorb salts from soil colloids is that the mycorrhizae increase the supply of exchangeable hydrogen ions, derived in part, at least, from carbonic acid. Their reasoning is based on the obvious fact that an electrostatic balance must be maintained in the culture medium. In other words, a mutual transfer of ions is involved in salt absorption by roots; for each cation absorbed by a root from a soil colloid an electrostatically equivalent anion must be yielded by the root surface.

Kessell and Stoate (1936) reported that difficulties were encountered in nurseries for the production of conifers in western Australia. It was found, however, that the trouble could be avoided if surface soil from older nurseries or the fruiting bodies of *Rhizopogon luteolus* were spread on. Treatment of this sort became standard practice. Rayner (1938) cited numerous examples from various parts of the British Empire of the successful use of inocula in growing exotics and in afforestation. Although the mycorrhizae problem is still far from solution in all details, it is now evident that mycorrhizal seedlings commonly exhibit greater growth and nutrient uptake than nonmycorrhizal

seedlings. This statement is particularly true of trees growing in soils having a low level of nutrients. The data presented in Table 15 were obtained by Mitchell, Finn, and Rosendahl (1937) and are typical of the results reported by Hatch (1936), White (1941), and McComb (1943). The theory now most favored to

TABLE 15. DRY WEIGHT AND MINERAL NUTRIENT ABSORPTION OF EASTERN WHITE PINE SEEDLINGS (Based on samples taken August 20 in the second growing season)

(After Mitchell *et al.*, 1937)

	Dry Weight, milligrams	Nitrogen Content, percentage	Phosphorus Content, percentage	Potassium Content, percentage
Seedlings with many mycorrhizae	337.2	1.6	0.21	0.63
Seedlings with very few mycorrhizae	180.6	1.2	0.07	0.45

account for increased growth and nutrient uptake is that the mycorrhizae raise the efficiency of the root system by greatly increasing the absorbing surface area.

Some fungi are rather specialized, developing only under certain conditions or on specific media; other fungi are able to grow under a wide variety of circumstances. Consequently, it is unsafe to generalize broadly as to the ecological conditions for the development of fungi as a whole.

Jensen (1931) remarked that, since the work of Ramann *et al.* (1899), it has been nearly a dogma in soil microbiology that "fungi prevail in acid, bacteria in neutral and alkaline soils." It is now recognized that occasionally fungi are abundant in neutral and alkaline soils. Evidently they will grow within a very wide range of reaction, from about pH 2.0 to 11.0. On the acid side fungi are much more tolerant than bacteria, but not on the alkaline side (Waksman, 1924; 1932: 246). The ratio between the number of bacteria and the number of fungi is influenced by the acidity of the soil; as the hydrogen-ion concentration decreases, the number of bacteria increases and the ratio becomes wider (Rippel, 1931: 260). Süchting (1929) investigated the development of eighteen species of forest soil fungi on media of varying pH; with one exception (*Mucor ramannianus*) all developed best within the range of pH 5.0-8.0.

Fungi are known to be able to modify the reaction of their media by the production of organic acids from carbohydrates, by the consumption of organic acids, and by the formation of ammonia from proteins (Waksman, 1932: 247). This fact was also recognized by Fehér (1933a: 158) and Fehér and Besenyei (1933b), who pointed out that curves of pH and numbers of soil fungi show fair agreement. Since the pH of the soil is strongly influenced by the activity of the fungi, the variations in pH may be regarded as an effect, not a cause, of variations in numbers of fungi.

In general, fungi are pronounced aerobes. The decrease in fungi with increasing depth below the soil surface was explained on this basis by Fehér and Besenyei (1933a). Fehér (1933a) found fungi relatively inactive at a depth of 40–50 cm., even in sandy soils.

Environmental conditions other than reaction and aeration also influence the development of fungi. There is reason to believe that the most favorable conditions for fungi and bacteria are similar (Süchting, 1929). Temperature and moisture conditions evidently have a dominant effect in determining the numbers of these organisms in forest soils. Fehér and Besenyei (1933a, 1933b) found a marked seasonal variation in the fungus population of European forest soils. The number of fungi reached its maximum in the summer or early fall months, and its minimum in winter. In general, the variations in number of fungi paralleled the variations in number of bacteria. As a rule, soil supporting coniferous forest stands contains more fungi than soil under deciduous forest stands (Fehér, 1933a).

PHYCOMYCETES

This group includes the Zygomycetes, the most important division of which is the order Mucorales. Representatives of the genera *Mucor*, *Zygorhynchus*, and *Rhizopus* are among the most common fungi of forest soils (Waksman, 1932: 241). A common *Zygorhynchus* was found by Waksman (1922) to occur abundantly in Sassafras and Lakewood soils in the Pine Barrens of New Jersey. In some places the fine, cottony mass of mycelia was holding the sand together in clumps. Powers and Bollen (1935) investigated a DeKalb forest soil from Pennsylvania and Aiken

and Olympic forest soils from Oregon; they found representatives of the genus *Mucor* among the most numerous molds. *Mucor ramannianus* was isolated by Melin (1934) from the *F* layer of a mixed conifer forest on Mount Desert Island, Maine. He regarded this species as characteristic of soils bearing conifer stands, since he isolated it frequently from such soils in Sweden. Species of *Mucor* are also important elements in the fungal flora of Hungarian soils (Fehér, 1933a: 167). On the other hand, Cobb (1932) found members of the Mucorales group scarce in both eastern hemlock and deciduous forest soils in New York. Svinhufvud (1937a) stated that in cold climates the soils are rich in *Mucor* and *Penicillium* species, whereas soils in warmer climates contain larger numbers of *Aspergilli*.

The Mucorales are, as a rule, unable to decompose cellulose and most hemicelluloses, but they decompose pectins (Waksman, 1932; Fehér and Besenyei, 1933a). Evidently little is known concerning the ecological requirements of fungi in the Zygomycetes group. It may be pointed out, however, that Sütting (1929) found that *Mucor ramannianus* developed best in highly acid conditions, that is, at pH 3.0.

ASCOMYCETES

The role played by members of this group in forest soils is not clear. In general, they are unable to decompose lignin material (Fehér and Besenyei, 1933a). Evidently they function in the decomposition of cellulose.

FUNGI IMPERFECTI

Members of the genera *Fusarium*, *Trichoderma*, *Aspergillus*, and *Penicillium*, which belong in the Hyphomycetes group, are among the most common soil fungi (Waksman, 1932: 241). Cobb (1932) reported that species of *Penicillium* (including *Citromyces*) and *Trichoderma* formed by far the largest part of the fungus population in eastern hemlock and in deciduous forest soil in New York. The *Penicillium-Citromyces* species were most numerous in the eastern hemlock soil; members of the genus *Aspergillus* were very scarce. The fungi were responsible for most of the decomposition in the eastern hemlock soil, but bac-

teria were the principal agents operating in the deciduous forest soil. In soil supporting loblolly pine (*Pinus taeda*) in Texas the dominant fungi belonged to the genus *Penicillium* and the closely related genus *Citromyces* (Morrow, 1932). *Penicillia* and *Aspergilli* were very numerous in the Pennsylvania and Oregon soils investigated by Powers and Bollen (1935). The importance of the Hyphomycetes, particularly the genera *Penicillium* and *Aspergillus*, in European forest soils has been emphasized by Fehér and Besenyei (1933a) and Fehér (1933a).

Heukelekian and Waksman (1925) investigated the carbon and nitrogen transformations in the decomposition of cellulose by two typical soil fungi, a *Trichoderma* and a *Penicillium*. It was found that the cellulose was completely decomposed, giving carbon dioxide as the only waste product. No intermediate products were left in the medium. Considerable of the carbon was reassimilated and built into protoplasm. Ammoniacal nitrogen proved more acceptable to the organisms than nitrate nitrogen. There is reason to believe that the Hyphomycetes may play a more important part than bacteria in the decomposition of cellulose in the soil.

BASIDIOMYCETES

The Basidiomycetes are a very important group of fungi in forest soils (Waksman *et al.*, 1928; Waksman, 1932). A number of the Agaricaceae, such as *Coprinus*, *Agaricus*, and *Lenzites*, are capable of developing in soils. Fehér and Besenyei (1933a) regarded the following families as the most important: Telephoraceae, Clavariaceae, Hydnaceae, and Polyporaceae.

The importance of the Basidiomycetes arises from the fact that representatives of the group are capable of decomposing both cellulose and lignin. Falck (1926) and Fehér (1935) believed that the destruction of woody cellulose is accomplished in large measure by the higher Basidiomycetes. Evidently the number of species which play a significant role in forest soils is relatively small.

ACTINOMYCES

Actinomyces, or ray fungi, resemble the filamentous fungi in that they produce an extensive unicellular, filamentous network.

They have a wide occurrence and in many soils are second only to bacteria in abundance.

In general, the occurrence of *Actinomyces* is dependent on about the same conditions as those which govern the development of bacteria. Since they are sensitive to excessive moisture, they occur in minimum numbers in water-logged soils. Dubos (1928) concluded that the activities of *Actinomyces* are limited to fairly dry soils. It appears that high acidity is unfavorable; Jensen (1930) worked in a number of Danish forests and found that acid soils (pH 4.6 or less) contained very small numbers of *Actinomyces*. Jensen found the largest numbers in soils having a pH of 6.8–8.0. Waksman (1932: 288) stated that the limiting reaction for the majority of forms is pH 4.8–5.0, but he recognized that some species may grow in media as acid as pH 3.0 or 4.0. In general, *Actinomyces* are regarded as scarce in soils having mor types of humus layers (Waksman and Starkey, 1931: 66). Eastern hemlock forest soil in New York, however, contained 215,050 *Actinomyces* per gram, and soil from a deciduous forest contained 555,550 (Cobb, 1932).

Actinomyces are important in the decomposition of soil organic matter, both nitrogenous and nonnitrogenous. Waksman and Diehm (1931) found *Actinomyces* even more active than fungi in the decomposition of cellulose. Mineralization of the organic matter of the soil is one of the functions of *Actinomyces*.

Statements that *Actinomyces* are capable of fixing atmospheric nitrogen must be regarded as unproved. Allison *et al.* (1934) investigated five common species and concluded that they cannot perform this function.

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Chapter 6

THE ORGANIC MATTER OF FOREST SOILS

The appearance of vegetation on the inorganic materials derived from rock weathering marks the beginning of a highly important phase in soil development. New substances, organic in nature, are added to the mineral material. This organic matter effects changes which have important physical, chemical, and biological results. Physical properties, such as soil color, structure, consistency, field capacity, volume weight, penetrability, permeability to water, and aeration, are modified by organic matter. Chemical effects of organic matter are evident in increased solubility of minerals, formation of various adsorption and chemical compounds with inorganic soil constituents, and acidity and buffer phenomena. Biological effects include supply of an energy source for soil organisms and liberation of nutrients for higher plants.

SOURCE OF SOIL ORGANIC MATTER

Soil organic matter is derived from the remains of relatively recent life, both plant and animal. In forests the principal sources are the leaves, stems, branches, roots, bark, fruits, and seeds of trees. The contribution made by shrubby and herbaceous vegetation is smaller, but not without significance. Animals also account for substantial quantities of valuable material, and the remains of microorganisms in the soil are a further source. Stepanoff (1929) reported that analyses of organic matter reaching the ground under a Scotch pine forest in Russia showed it to have the following composition: branches, 8.6; pine needles, 49.4; bark and cones, 29.2; and leaves of undergrowth, 12.8 per cent.

The weight of dry organic matter returned to the soil annually by stands of different species growing in the same general environment appears to be about the same (Ebermayer, 1876, 1890). Ebermayer gave 2971 lb. per acre as the average amount of dry organic matter, mostly leaves, contributed annually to the soil

by European beech (*Fagus sylvatica*) stands. The corresponding value for Norway spruce was 2682 and for Scotch pine 2842. The age of the stands investigated by Ebermayer varied from 25 to 120 years. The average fall of litter in jack and Norway pine (*Pinus resinosa*) forests in northern Minnesota during a 12-month period was found to be 1738 lb. per acre, oven-dry weight (Alway and Zon, 1930). Sims (1932) reported that the annual leaf fall in a pine-oak stand on a south slope in the Bent Creek Experimental Forest in North Carolina amounted to 2600-3100 lb. per acre, oven-dry weight. According to Heyward and Barnette (1936), the annual needle fall in second-growth slash (*Pinus caribaea*) and longleaf pine in Florida was approximately 2400-3500 lb. per acre, oven-dry weight. Measurements made by Morgan and Lunt (1932) in 27-year-old stands of eastern white pine and Norway pine in southern New England indicated the fall of litter to amount to 2000-3000 lb. per acre annually. Chandler (1941) has reported values for hardwood leaf litter in New York ranging from 2425-3020 lb. of dry matter per acre annually.

In addition to the litter which annually reaches the ground under forests, further substantial amounts of organic matter in the form of slash are added at irregular periods as a result of intermediate and reproduction cuttings. The amount of slash resulting from cutting operations in Douglas fir stands on the Pacific coast has been found to average about 24,000 cu. ft. per acre (Isaac and Hopkins, 1937). It is estimated that this quantity constitutes roughly 720,000 lb. of dry matter per acre.¹

CHEMICAL COMPOSITION OF FRESH PLANT MATERIAL

Fresh plant material is composed of a large number of complex chemical compounds, most of which are nonnitrogenous. Composition varies considerably in different plant parts and in different species. Likewise, the age of the material and the environmental conditions under which it developed have an influence on the constituents present. The more important plant substances, as grouped by Waksman (1932), may be indicated as follows:

1. Sugars, starches, and simple carbohydrates, most of which are soluble in cold or hot water.

¹ Based on the assumption that the material weighs 30 lb. per cubic foot.

2. Pentosans, pectins, and other hemicelluloses, such as galactans and mannans, readily hydrolyzed by dilute acids.
3. True cellulose.
4. Lignins and tannins.
5. Fats, waxes, oils, sterols, and fatty acids.
6. Proteins and their derivatives.
7. Mineral constituents.

Mature plant residues, which are composed largely of cell-wall substances, consist principally of the following chemical compounds (Waksman, 1938: 95):

	Percentage
Cellulose	20-50
Hemicelluloses (pentosans, hexosans, and polyuronides)	10-28
Lignins	10-30
Tannins, coloring substances, cutins, suberins, fats, waxes	1-8
Proteins	1-15

Calcium, potassium, magnesium, phosphorus, and silicon are relatively abundant mineral elements in tree leaves. In addition, there are many others which are less abundant. Analysis of leaves of a hickory tree growing in soil formed from a vein of pegmatite in Virginia resulted in the identification of more than 50 chemical elements, including 11 rare-earth elements (Robinson *et al.*, 1938). Later investigations by Robinson (1943), employing some 70 samples of hickory leaves from various localities, confirmed the earlier results. *Carya* spp. appear to absorb larger amounts of the rare-earth elements than such other trees as *Pinus*, *Liriodendron*, *Cornus*, *Quercus*, *Betula*, *Acer*, and *Fraxinus*. Copper, manganese, and zinc have been found to be normal constituents in the leaves of 23 different hardwood trees in Kentucky (McHargue and Roy, 1932). Wolff (1880) and Kleinstück (1928) showed that considerable manganese may occur in the leaves and bark of certain hardwoods and conifers.

COMPOSITION OF VARIOUS PARTS OF FOREST TREES

Various parts of a given tree vary greatly in both mineral and organic constituents. The variations in mineral constituents are illustrated by the data of Henry (1908), which appear in Tables 16 and 17. This material was obtained in the month of June

from trees in a 37-year-old stand growing on relatively uniform soil in Haye Forest, France.

TABLE 16. CONTENT OF PURE ASH IN PARTS OF
VARIOUS TREES (Values represent percentages based on dry matter)
(After Henry, 1908)

Tree Part	Tree Species			
	<i>Populus tremula</i>	<i>Fagus sylvatica</i>	<i>Fraxinus excelsior</i>	<i>Quercus robur</i>
Stem wood	0.398	0.355	0.361	0.311
Branches	2.275	1.590	1.831	2.426
Bark	3.334	5.857	4.114	5.575
Leaves	8.865	5.142	7.001	4.508

TABLE 17. COMPOSITION OF THE ASH FROM PARTS OF
VARIOUS TREES (Values represent percentages based on ash content)
(After Henry, 1908)

Chemical Con- stituent	Tree Part	Tree Species			
		<i>Populus tremula</i>	<i>Fagus sylvatica</i>	<i>Fraxinus excelsior</i>	<i>Quercus robur</i>
K ₂ O	Stem	11.8	14.5	13.2	9.8
	Branches	9.0	11.8	17.6	14.2
	Bark	7.7	5.1	8.3	2.9
	Leaves	18.4	21.8	18.7	22.4
CaO	Stem	71.2	60.2	62.1	76.2
	Branches	77.8	66.5	64.4	69.2
	Bark	72.8	83.4	80.2	87.9
	Leaves	49.6	44.3	39.4	47.0
MgO	Stem	3.8	4.5	5.8	3.6
	Branches	5.3	5.3	3.1	6.3
	Bark	7.1	3.6	2.3	2.0
	Leaves	4.0	7.3	8.1	2.9
Fe ₂ O ₃	Stem	1.2	2.3	1.8	1.5
	Branches	1.1	1.6	1.6	0.5
	Bark	3.0	0.7	1.2	1.0
	Leaves	2.0	2.3	1.1	3.1
P ₂ O ₅	Stem	4.4	2.8	6.8	3.6
	Branches	3.1	6.1	8.4	4.6
	Bark	3.1	2.1	3.8	2.9
	Leaves	8.8	7.8	22.6	12.4
SiO ₂	Stem	2.7	10.0	2.2	2.3
	Branches	2.2	5.6	1.8	1.6
	Bark	2.2	3.7	1.4	0.8
	Leaves	7.5	10.5	2.6	5.2

The relative ash content of tree organs is in general as follows: leaves > bark > branches > stem wood. Sometimes the bark contains more ash material than the leaves. As indicated in Tables 16 and 17, branches generally have a higher ash content than stem wood. This fact has also been recognized by Schütze (1871), Morosov (1928: 98), and Dengler (1935: 151). It appears that the bark of trees commonly contains more CaO and less MgO, P₂O₅, and K₂O than the wood (White, 1889; Henry, 1908).

The ash content of domestic woods generally falls within the limits of about 0.2–1.0 per cent (White, 1889; Schorger, 1926: 50; Kurth, 1944: 434). Occasionally maxima of about 3.0 or 4.0 per cent are reported. Sapwood commonly contains a greater amount of ash than does heartwood, but exceptions are known to occur. The percentage of CaO in the ash of wood is, in general, relatively high; it rarely drops below 20 and may amount to 60–70 or more (Hawley and Wise, 1926: 119–120). Quite frequently the percentage of K₂O in wood ash ranges between 10 and 30, and the content of MgO between 5 and 10. Wolff (1880) concluded that, in general, the wood and bark of conifers contain less ash than those of hardwood trees. He reported that the bark of conifers generally contains 1.5–3.0 per cent ash, whereas that of hardwoods usually contains 3.0–9.0 per cent. Ramann (1890: 52) reported the nitrogen content of wood as 0.1–0.5 per cent. More recently Kalniņš and Liepiņš (1938) found the nitrogen content of Scotch pine wood to be 0.12–0.16 per cent.

In forests under management tree leaves and branches contribute much larger quantities of organic debris and mineral matter to the soil than does stem wood. Tree leaves are by far the largest single source of forest soil organic matter, and their annual contribution of mineral elements to the soil greatly exceeds that of all other tree parts combined. For this reason the student of forest soils is particularly concerned with both the amount of leaf litter and its chemical composition.

Artificial fertilization of forest trees in America is, under all but exceptional circumstances, not economically feasible at the present time. Therefore the trees must rely solely on the natural supply of nutrients for their requirements. In the majority of forest soils a substantial part of the currently available nutrients comes from the litter which annually reaches the ground. If the leaves contain a large amount of a certain element, the content

of this constituent in the surface soil layers will tend to be maintained at a level higher than that of the horizons below. Thus, there is a distinct tendency in forest soils for the nutrient capital to be concentrated in the uppermost layers or horizons. For this reason forests are commonly credited with the ability to restore fertility in soils worn out by unsound agricultural use. Nutrient elements which would otherwise be lost in drainage water are absorbed by the deeply penetrating tree roots and returned to the soil surface in the annual leaf fall.

The nutrient content of forest tree leaves varies according to the species, age of the leaves, and composition of the soil. Comprehension of the effects of these factors is essential to an understanding of the influence of trees on the humus layers of forest soils.

COMPOSITION OF FOREST TREE LEAVES AS INFLUENCED BY SPECIES

Tree species differ greatly in the relative and absolute amounts of nutrient elements absorbed from the soil. Consequently, the composition of the leaves of various species is different, even when the trees are growing in the same habitat and drawing their nutrients from the same soil solution. Differences in chemical composition account for the fact that the litter of some species is more valuable as fertilizer than that of others. Chemical composition of the organic debris also influences the rate and nature of decomposition. Representative data for twenty-four tree species from various sections of the United States are presented in Table 18.

The calcium content of forest tree leaves has received particular attention from soil scientists. Interest in this element is based on the fact that it exerts a remarkable influence on soil properties. A high content of calcium in the litter increases the amount of exchangeable calcium in the soil, reduces soil acidity, favors development of mull humus layers, and stimulates the activity of the soil fauna and flora.

In spite of the fact that variations of chemical composition occur within a tree species, depending primarily on environmental conditions, it is clear that specific differences also exist. This fact is illustrated by the data in Table 19. Inspection of Table 19

TABLE 18. NUTRIENT ELEMENT CONTENT OF FRESHLY
FALLEN LEAF LITTER FROM COMMON FOREST TREES OF
THE UNITED STATES* (Values represent the element, not oxide,
expressed as a percentage of dry weight)

Species	Cal- cium	Magne- sium	Potas- sium	Phos- phorus	Nitro- gen	Ash
<i>Abies balsamea</i>	1.12	0.16	0.12	0.09	1.25	3.08
<i>Acer rubrum</i>	1.29	0.40	0.40	0.09	0.52	10.97
<i>A. saccharum</i>	1.81	0.24	0.75	0.11	0.67	11.85
<i>Betula lenta</i>	1.65	0.28	0.75	0.17	0.72	
<i>Carya ovata</i>	2.09		1.18	0.16	0.91	9.59
<i>C. cordiformis</i>	3.41	0.55	0.44	0.12	0.67	
<i>Fagus grandifolia</i>	0.99	0.22	0.65	0.10	0.67	7.37
<i>Fraxinus americana</i>	2.37	0.27	0.54	0.15	0.63	10.26
<i>Liriodendron tulipifera</i>	2.56	0.45	0.95	0.11	0.51	10.67
<i>Magnolia acuminata</i>	1.71	0.29	0.76	0.28	0.58	
<i>Picea abies</i>	1.96	0.23	0.39	0.09	1.02	
<i>P. rubens</i>	0.79	0.20	0.35	0.10	0.89	
<i>Pinus banksiana</i>	0.61		0.16	0.04	0.58	4.15
<i>P. echinata</i>	0.75				1.00	3.50
<i>P. resinosa</i>	0.76	0.18	0.30	0.07	0.68	4.27
<i>P. strobus</i>	0.72	0.19	0.19	0.07	0.70	4.33
<i>P. taeda</i>	0.55				0.80	3.01
<i>Populus tremuloides</i>	2.11	0.23	0.47	0.10	0.73	
<i>Prunus serotina</i>	2.37	0.42	0.55	0.18	0.60	
<i>Quercus alba</i>	1.36	0.24	0.52	0.13	0.65	5.71
<i>Q. borealis</i>	1.38	0.25	0.65	0.11	0.64	6.84
<i>Thuja occidentalis</i>	2.16	0.15	0.25	0.04	0.60	
<i>Tilia americana</i>	3.84	0.77	0.52	0.12	1.01	15.16
<i>Tsuga canadensis</i>	0.68	0.14	0.27	0.07	1.05	

* Average values derived from data published by Melin (1930), Alway, Kittredge, and Methley (1933), Alway, Maki, and Methley (1934), Lunt (1933, 1935), Coile (1937), and Chandler (1941, 1944).

indicates that the leaves of a given species, although obtained from widely separated localities, tend to be fairly consistent in calcium content. Data obtained at Cornell University indicate that individuals of a given species, growing on the same site, may exhibit fully as great variability in chemical composition as individuals growing in widely separated regions.

The analytical data in Tables 18 and 19, supplemented by additional determinations from various sources, permit recognition of three groups of tree species based on the calcium content of the foliage. The first group includes trees whose mature leaves usually contain more than 2 per cent calcium: *Tilia amer-*

TABLE 19. A COMPARISON OF THE CALCIUM CONTENT OF FOREST TREE LEAVES AS REPORTED BY VARIOUS INVESTIGATORS (Values represent the element, expressed as a percentage of dry weight)

Species	Calcium Content, percentage				Average
<i>Pinus strobus</i>	0.57*	0.60†	0.55‡	1.03§	0.69
<i>Fagus grandifolia</i>	0.97*	0.75	0.87¶	1.04**	0.91
<i>Quercus alba</i>	1.13*	1.36¶	1.37**	1.10††	1.24
<i>Q. borealis</i>	1.29*	1.38**	1.21	1.04‡‡	1.23
<i>Acer saccharum</i>	1.80*	1.59§§	1.67**	1.75	1.70
<i>Liriodendron tulipifera</i>	2.56**	3.24	3.03‡‡	3.00	2.96
<i>Tilia americana</i>	3.23**	3.62‡‡	2.81	4.42¶¶	3.52

* Broadfoot and Pierre (1939).

† Chandler (1944).

‡ Plice (1934).

§ Alway, Kittredge, and Methley (1933).

|| Chandler (1939).

¶ Lunt (1933).

** Chandler (1941).

†† Mitchell (1936).

‡‡ Chandler (1937).

§§ Lunt (1935).

|||| Coile (1937).

¶¶ Alway, Maki, and Methley (1934).

icana, *Juniperus virginiana*, *Carya cordiformis*, *C. ovata*, *Liriodendron tulipifera*, *Fraxinus americana*, *Prunus serotina*, *Thuja occidentalis*, *Populus tremuloides*, *Celtis occidentalis*, *Ulmus americana*, and *Cornus florida*. In the second group are species whose foliage generally contains between 1 and 2 per cent calcium: *Picea abies*, *Acer saccharum*, *A. rubrum*, *Magnolia acuminata*, *Betula lenta*, *B. lutea*, *Quercus borealis*, *Q. alba*, and *Abies balsamea*. The third group includes species whose leaves usually contain less than 1 per cent calcium: *Fagus grandifolia*, *Picea rubens*, *Pinus echinata*, *P. resinosa*, *P. strobus*, *P. banksiana*, *P. taeda*, *P. sylvestris*, and *Tsuga canadensis*. These listings should be regarded as tentative and not final; some shifting of species from one group to another may become necessary as information increases. Data concerning the calcium content of leaves of trees in the western United States are insufficient to warrant their classification in the above groups at this time. Available information indicates, however, that the calcium content of the pines is lower than that of the firs and cedars.

In general, the leaves of hardwoods contain more calcium than leaves of conifers, yet it should be recognized that there are some exceptions. For example, foliage of eastern red cedar and northern white cedar generally contains far more calcium than does that of American beech.

Although the data for magnesium, potassium, and phosphorus content of leaves show a variability between species amounting to several hundred per cent, enough studies have not been conducted to demonstrate that the values are representative. It is evident, however, that the content of these elements tends to be higher in hardwood leaves than in conifer leaves. There appears to be a tendency for leaves of conifers to contain more nitrogen than those of hardwoods. Five species are indicated in Table 18 as having leaves with a nitrogen content of 1.00 per cent or more; four of these are conifers, and only one is a hardwood. Classification of tree species according to their content of magnesium, potassium, phosphorus, and nitrogen may well be deferred until additional analyses are available.

The acidity of leaves of different tree species varies considerably and undoubtedly influences the acidity of the soil. Hesselman (1926) suggested that characteristic differences in the pH values of leaf material exist between species and groups of species, but his data indicate that within the same locality the pH of birch leaves may vary from 5.3 to 6.1 and that values for European beech may vary from 5.3 to 6.6. Plice (1934) investigated leaf material from twenty-seven species of hardwoods and conifers and found that most of them come within the narrow range of pH 4.1-4.5. The hydrogen-ion concentration of tree leaves is to some extent influenced by the reaction of the soil (Olsen, 1932). In general, the leaves of conifers are more acid than those of hardwoods. Data concerning the pH of leaves of several species are presented in Table 20.

The acid buffer and basic buffer² content of tree leaves has been regarded by Hesselman (1926) as fairly characteristic for different species. He recognized the following groups of plants, classified on the acid and basic buffer content of their leaves:

1. Those having high acid and low basic buffer content, for example, many conifers.

² A discussion of buffer action will be found in Chapter 10.

2. Those having very high acid and fairly high basic buffer content, for example, various hardwood species and many herbaceous plants characteristic of deciduous forests.
3. Those having very low acid and very high basic buffer content.
4. Those having high acid and high basic buffer content.
5. Those having especially low buffer content.

TABLE 20. THE ACIDITY OF FOREST TREE LEAVES

Species	pH Classes			
	3.0-3.9	4.0-4.9	5.0-5.9	6.0-6.9
<i>Abies balsamea</i>	*	†		
<i>Acer rubrum</i>		*, †		
<i>A. saccharum</i>		*, †		
<i>Betula papyrifera</i>		*	*	
<i>B. populifolia</i>		*		
<i>Carya glabra</i>			†	
<i>Fagus grandifolia</i>		†	*, †	*
<i>Fraxinus americana</i>		†	†	*
<i>Larix decidua</i>		*, †		
<i>Picea rubens</i>		*		
<i>Pinus banksiana</i>	†			
<i>P. resinosa</i>		†		
<i>P. rigida</i>		*, †		
<i>P. strobus</i>		*, †		
<i>Populus grandidentata</i>			*	
<i>Quercus alba</i>		†		
<i>Q. borealis</i>		†		
<i>Thuja occidentalis</i>		†		
<i>Tsuga canadensis</i>	†			
* Melin (1930).				
† Plice (1934).				

Hardwood leaves appear to contain a predominance of basic buffer substances, and conifers a predominance of acid buffer substances. Plice (1934) investigated the antacid buffer capacity of leaf material from 27 tree species, of which 13 were conifers and 14 were hardwoods. Five grams of *Magnolia acuminata* leaves inactivated 3.98 milligram equivalents (m.e.) of hydrogen ion out of 4.00 added, whereas only 0.9 m.e. were inactivated by one of the 5.0-gram samples of *Tsuga canadensis* leaves. The 27 species employed may be separated into two groups, based on the milligram equivalents of hydrogen ion inactivated by 5.0-gram samples of leaves. In group 1, with 3.0 or more m.e. of hydrogen

ion inactivated, out of 4.0 m.e. of hydrogen ion added, are 3 conifers and 13 hardwoods; in group 2, with less than 3.0 m.e. inactivated, are 10 conifers and 1 hardwood.

The composition of the leaves of shrubs and herbs which occur in forest stands has not received much attention in America. Ebermayer (1876: 115, 124) presented data on various European species which indicated that they may be rich in phosphorus and potash. Similar results were reported by Plice (1934) for a few American species. Litter of *Oxalis acetosella* was found to contain 2.2 per cent potash, 0.8 per cent phosphoric acid, 1.0 per cent magnesia, and 2.5 per cent nitrogen.

COMPOSITION OF FOREST TREE LEAVES AS INFLUENCED BY AGE

The composition of forest tree leaves varies with age. This variation is indicated by the data of Alway *et al.* (1934), as presented in Table 21. These data, based on material collected in

TABLE 21. CHEMICAL COMPOSITION OF FOREST TREE LEAVES COLLECTED IN DIFFERENT SEASONS (Values represent the element, expressed as a percentage of dry weight, and averaged for nine hardwood species)

Date of Collection	Ash	Ca	Mg	P	K	S	N
June 1	6.59	1.20	0.37	0.29	1.49	0.20	2.99
July 1	6.89	1.49	0.47	0.25	1.16	0.18	2.36
August 1	8.22	1.92	0.49	0.22	1.09	0.17	2.24
September 1	10.24	2.23	0.54	0.22	0.99	0.18	1.85
October 11-16*	11.32	2.85	0.57	0.19	0.41	0.14	0.80

* Leaves freshly fallen.

Minnesota, are typical of results reported by other investigators (Rissmüller, 1874; Serex, 1917; McHargue and Roy, 1932; Mitchell, 1936). In essentially all instances, and for all species that have been investigated, the percentage (based on dry matter) of phosphorus, potassium, and nitrogen in the foliage decreases as the growing season progresses. On the other hand, the percentage of ash, silicon, magnesium, and calcium apparently increases as the season advances. The pronounced decrease in the percentage of phosphorus, potassium, and nitrogen (based on dry matter) between September 1 and the middle of October is of interest.

This decrease has commonly been attributed to migration of these elements from the leaves back into the branches and stem. It is also thought that toward the end of the growing season rainfall may cause some leaching from the leaves. Evidently this leaching does not occur before yellowing of the foliage (Mitchell, 1936).

It should be recognized that a decrease in the percentage (based on dry matter) of an element does not necessarily mean a decrease in the absolute amount in the leaves. Actually the reverse may be true. Mitchell (1936) has worked on this problem; his data for oak leaves are presented in Table 22. The

TABLE 22. NUTRIENT CONTENT OF FOREST TREE LEAVES COLLECTED AT VARIOUS TIMES DURING THE GROWING SEASON (Values are averages for *Quercus alba*, *Q. borealis*, and *Q. montana*, expressed in milligrams per average leaf)

Date of Sampling	Dry Weight of Average				
	Leaf	N	P	K	Ca
May 27	192	7.19	0.66	2.82	1.04
June 26	386	10.65	0.67	4.66	2.52
July 26	485	12.78	0.73	5.03	3.66
August 26	530	13.11	0.80	5.31	4.82
September 25	551	13.25	0.82	5.51	5.25

trend indicated in this table is characteristic of the species which have been investigated. As the growing season progresses, the dry weight of the individual leaves increases, as does the actual content of nutrient elements. However, the dry-matter content of the leaves increases more rapidly than do the nutrient elements, with the result that there is an apparent decrease in these elements when they are expressed as a percentage of the dry weight. After yellowing of the leaves begins in the autumn, there is an actual decrease in the content of phosphorus, potassium, and nitrogen, which is due to migration of these elements from the leaves back to the branches. This decrease is not indicated by the data in Table 22, because the leaves were still green on September 25, the last date of sampling.

The calcium content of eastern white pine needles, as influenced by age, is presented graphically in Figure 19 (Chandler, 1939). It will be observed that there is a progressive increase of calcium content in new needles from June 1 to October 1. This increase

continues during the growing season of the second and third years that the leaves remain on the trees. Evidently no appreciable changes in calcium content occur during the winter periods as long as the leaves remain green. Similar seasonal trends have been secured for the calcium content of foliage of eastern red cedar. It is probable that a large proportion of the calcium which

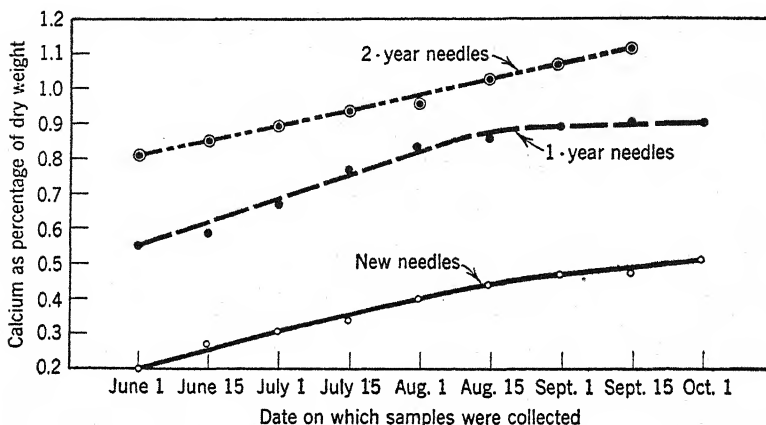


FIG. 19. Calcium content of new, 1-year, and 2-year eastern white pine needles throughout the growing season, expressed as a percentage of dry weight.

accumulates in tree foliage occurs as the oxalate; in this form it is relatively inert (Chandler, 1937).

The trend of curves showing the changing composition of tree leaves as the growing season advances indicates that modifications during the latter part of the season (but before yellowing of the foliage) are relatively small in comparison to those occurring in the early and middle parts of the growing season. Therefore, if green leaves are to be analyzed for nitrogen and mineral nutrient content, it seems advisable to collect the samples during the 2- or 3-week period before yellowing.

Henry (1908: 17) presented data which indicated that the dead bark of Norway spruce contained considerably less K_2O and more SiO_2 than living bark. The content of P_2O_5 and CaO was approximately the same in both cases. Quite often the average ash content of the total wood of an old tree is lower than that of

a young tree of the same species (Hawley and Wise, 1926: 119-120).

The organic compounds present in tree leaves also change with advancing age. Data obtained by Waksman *et al.* (1928) indicated a lower content of water-soluble and ether-soluble constituents in mature oak leaves than in younger leaves. The content of hemicelluloses, celluloses, and lignins was highest in the mature leaves.

COMPOSITION OF FOREST TREE LEAVES AS INFLUENCED BY ENVIRONMENT

Among environmental factors influencing the composition of forest tree leaves the chemical properties of soil are outstanding. This effect is more pronounced for some elements than others. The nitrogen, potassium, and even phosphorus levels in the soil

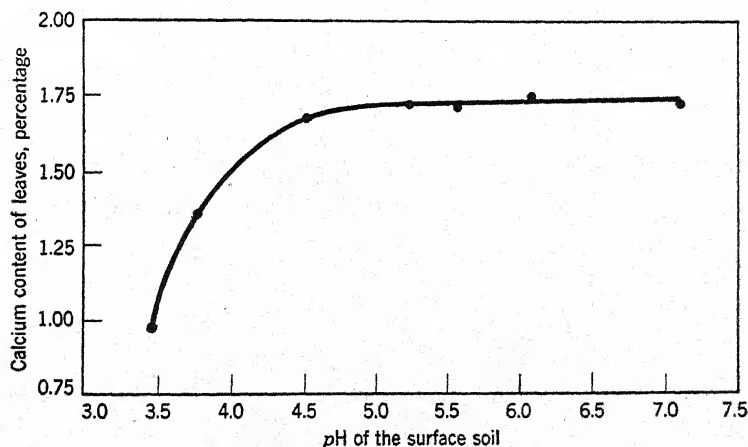


FIG. 20. Relationship between pH of the surface soil and average calcium content of leaves from sugar maple trees on various sites in New York.

seem to influence the nutrient content of tree leaves of a given species to a greater extent than do calcium and magnesium. This statement applies particularly to the range of nutrient levels normally encountered in forest soils. The relationship between soil composition and calcium content of tree leaves is illustrated

by data for sugar maple, obtained by Chandler (1941) and presented in Figure 20. Leaves of sugar maple trees growing in soils with a pH range of 4.5–7.0 contain about the same amount of calcium. It should be noted, however, that, when the soils are extremely acid and have a base-saturation percentage below 30, the calcium content of the foliage drops. Conversely, an unusually high concentration of calcium in the soil may be reflected in a rise in the content of the element in the foliage.

Although the calcium content of leaves of a given species does not seem to be appreciably influenced by variations within the range of calcium content usually encountered in soils, it should be recognized that tree litter on calcareous soils generally contains more calcium than that on acid soils. This phenomenon is largely explained by the occurrence on calcareous soils of tree species that normally have a high content of calcium in their leaves. Figure 21, taken from the work of Chandler (1941),

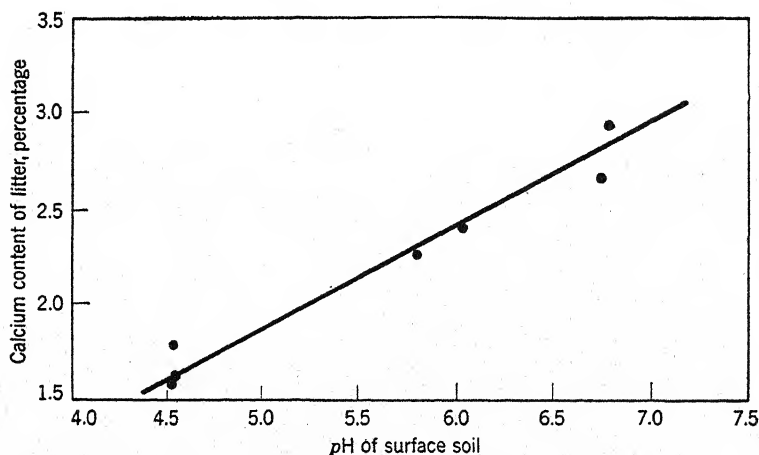


FIG. 21. Relationship between pH of the surface soil and average calcium content of tree litter on various sites in New York.

shows that in representative samples of mature, freshly fallen leaves, collected without regard to species, the percentage calcium content increased from about 1.7 to 2.7 as the pH of the soil increased from 4.53 to 6.76. The average percentage calcium content of the leaves of individual species, however, was 2.02 for the acid soils and 2.01 for soils high in exchangeable calcium.

A quite different relationship exists in regard to nitrogen. There seems to be a rather direct correlation between the level of so-called available nitrogen in soil and the content of this element in mature foliage. In view of the difficulties encountered in determining available nitrogen in forest soils, the relationship may be most readily demonstrated by applying nitrogen in available form to the soil and then analyzing the leaves for total nitrogen. Data reported by Mitchell and Chandler (1939) concerning nitrogen uptake by red maple are presented in Figure 22.

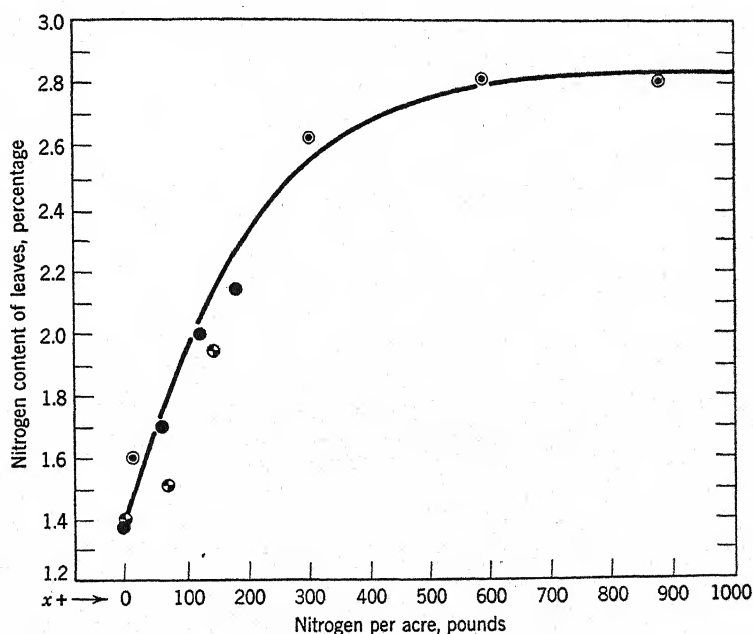


FIG. 22. The relationship between the nitrogen supply and the nitrogen content of leaves of red maple trees growing in even-aged stands of mixed hardwoods. Courtesy, Mitchell and Chandler (1939).

The nitrogen fertilizer which was applied consisted of equal parts of ammonium sulfate and sodium nitrate. In all species investigated, an increase in the nitrogen content of the leaves resulted from the application of nitrogenous fertilizers.

The phosphorus content of forest tree leaves is relatively low and does not appear to vary greatly in trees growing on different

sites. However, Mitchell and Finn (1935) have demonstrated that the content of this element in northern red oak or red maple leaves may be approximately doubled when phosphatic fertilizers are applied to forest soils. Additional investigations are necessary to clarify the relationship between phosphorus levels in forest soils and the phosphorus content of tree foliage.

The concentration of potassium in the leaves of forest trees has likewise not received adequate attention. Unpublished data secured at Cornell University demonstrate that the content of this element in tree foliage may be greatly increased by application of fertilizer. Results for different sites have proved rather variable, and no definite relationships have as yet been established.

Information concerning the relation between magnesium levels in forest soils and in tree foliage is scanty. Wherry (1932) showed that leaves of pitch pine (*Pinus rigida*) and blackjack oak (*Quercus marilandica*) growing in soils derived from serpentine contained less ash and potash but more magnesium than leaves of the same trees growing in areas lacking serpentine. It should be recognized, however, that the nutrient supply in soils derived from serpentine tends to be unbalanced, and that a high magnesium content in tree leaves under these conditions does not necessarily indicate any close relation between the magnesium content of normal forest soils and the abundance of the element in forest tree foliage.

AMOUNT OF NUTRIENTS RETURNED TO THE SOIL ANNUALLY BY FOREST TREE LEAVES

The amount and chemical composition of leaf litter which annually falls to the ground in forest stands have been considered in the foregoing pages. From data such as have been presented it is possible to calculate in a general way the amount of nutrients deposited annually on the soil under forest stands. Calculations of this kind, in spite of their limitations, are useful, for they contribute to an appreciation of the fertilizer value of leaf litter from various species and groups of species. Furthermore, they indicate in a broad way the amounts of nutrients taken up and the annual turnover of the nutrient capital.

In Table 23 are presented average values representing the

amounts of nutrients returned to the soil annually by hardwoods and conifers. These data, based on the work of Chandler (1941, 1944) in New York, indicate that hardwood species as a group return approximately twice as much calcium, magnesium, potassium, and phosphorus to the soil annually as do conifer species as a group. On the other hand, more nitrogen is returned to the soil under conifers than to that under hardwoods.

TABLE 23. AMOUNT OF NUTRIENTS RETURNED TO THE SOIL ANNUALLY BY FOREST TREE LEAVES (Values represent pounds per acre)

	Ca	Mg	K	P	N
Hardwoods	65.6	9.2	13.5	3.3	16.6
Conifers	26.5	4.5	6.5	1.8	23.6

In order to avoid unwarranted generalizations it is essential that the influence of species, age, and environment (especially the chemical properties of the soil) on the composition of tree leaves be kept in mind. Thus, although the leaf litter of conifers *as a group* is much lower in mineral nutrients than that of hardwoods *as a group*, it is a fact that the litter of certain conifers, such as Norway spruce and northern white cedar, contains more calcium than that of certain hardwoods, such as white oak and American beech. Failure to recognize species differences has contributed to the unsound generalization that hardwoods are "good" and conifers "bad" from the standpoint of soil fertility.

DECOMPOSITION OF PLANT RESIDUES AND DEVELOPMENT OF HUMUS

It is evident that organic debris contains substantial amounts of the nutrient elements needed by plants. However, the organic residues must be decomposed before the contained nutrients are again available for use. Waksman (1932: 676) expressed the opinion that the major problems of forest soil fertility concern the rapidity and nature of decomposition of the plant debris reaching the soil. Furthermore, decomposition is necessary if the fresh organic debris is to exert its most favorable physical, chemical, and biochemical influences.

The material which develops as a result of decomposition of organic matter is called *humus*. Waksman (1938) has defined

humus as a complex aggregate of brown to dark-colored amorphous substances, which have originated during the decomposition of plant and animal residues by microorganisms, under aerobic and anaerobic conditions, usually in soils, composts, peat bogs, and water basins. Many workers, particularly those concerned with the agronomic use of soils, prefer to define humus as the well-decomposed, more or less stable portion of the organic matter in soils. Strict adherence to this definition, however, is impossible in the field of forest soils if the generally accepted concept of humus layer types is employed. If the term *humus* is restricted to the well-decomposed, more or less stable portions of the organic matter in soils, then material constituting the *F* and *H* layers of fibrous mor (a humus layer type) is not humus. In view of this inconsistency a somewhat broader definition of humus is desirable. Consequently, in this book humus will be regarded as the plant and animal residues of the soil, litter excluded, which are undergoing evident decomposition.

Fresh organic debris reaching the soil is subjected to the influence of various agents such as water, air, plant enzymes, animals, and microorganisms. The first three of these agents bring about the oxidation of certain compounds and the hydrolysis of others; they also play a part in the interaction among various substances, such as the tannins, proteins, and lignins (Waksman, 1938: 120). As a result of animal activity the organic debris is reduced in amount, changed in chemical composition, and mixed with inorganic soil material. Earthworms of the genera *Lumbricus*, *Allolobophora*, and *Eisenia*, and various soil arthropods contribute to the breakdown of organic debris. However, the most important changes in organic debris are probably effected by the microorganisms of the soil.

The attack on organic debris by microorganisms leads to chemical changes as a result of oxidation, hydrolysis, reduction, and condensation. It has been commonly assumed that the decomposition of organic debris under aerobic conditions consists in a chemical simplification of the various complex compounds, finally leading to their transformation to carbon dioxide, ammonia, water, and mineral elements (Waksman, 1932: 676). It may be pointed out, however, that even under aerobic conditions, an attack on plant remains by microorganisms seldom results directly in complete decomposition. The readily decomposed

constituents are rather quickly broken down, partly to intermediate products and partly to end products. The intermediate products may undergo further alteration; the end products may either be utilized by microorganisms or, if in excess of their needs, liberated. The constituents more resistant to decomposition disappear only gradually; to a considerable extent they form the humus of soils. In the course of time the original plant material is for the most part transformed to various end products and to synthesized substances in the form of microbial cells.

Various theories, some of them purely chemical, have been advanced to explain the formation of humus. In recent years greater importance has been attached to the activities of microorganisms, as agents of both decomposition and synthesis, in the formation of humus. Waksman (1938: 107) has summarized the situation as follows:

It was found that in the decomposition of plant and animal residues, the carbohydrates tend to disappear rapidly and the lignins accumulate. The former serve as sources of energy for the microorganisms bringing about the decomposition, with the result that considerable cell substance, consisting largely of proteins and certain hemicelluloses, is synthesized; these compounds, together with the modified lignins of the plant residues, form the bulk of the constituents of humus. According to this [microbiological] theory, the carbohydrates, proteins, and lignins are associated with the formation of humus, directly or indirectly.

Sugars and other *water-soluble substances* are the first to be decomposed. Part may be completely oxidized to carbon dioxide and water, and part may be incompletely decomposed with the production of various products, such as organic acids (citric, oxalic, fumaric, glucuronic, succinic, lactic, butyric, acetic, formic, propionic, valeric, pyruvic). *Starches* are also decomposed rapidly, the processes being similar to those in the breakdown of sugars. *Pentosans* are decomposed rapidly, possibly more rapidly than any other group of water-insoluble complexes in fresh material (Waksman, 1929). *Hemicelluloses* vary in chemical composition and differ greatly in the ease with which they are decomposed by microorganisms. Some of the hemicelluloses are very resistant to rapid decomposition and may play an important part in the formation of humus. On hydrolysis hemicelluloses are changed to simple sugars, hexoses, or pentoses.

Cellulose is a very important constituent of plant residues

reaching the soil and represents a major source of energy for the most active groups of microorganisms. Among the fungi known to decompose cellulose are a considerable number of Ascomycetes, Basidiomycetes, and Fungi Imperfecti. In decomposing cellulose, fungi require about 1 part of nitrogen for every 30-50 parts of cellulose decomposed. For every 100 units of cellulose decomposed, fungi synthesize 20-30 units of cell substance, which contain 1.5-2.5 units of nitrogen. Thus the amount of available nitrogen frequently becomes a limiting factor in the decomposition of cellulose. Both aerobic and anaerobic bacteria also decompose cellulose. As a result of its decomposition, various products, in addition to cell substance, are produced, for example, hemicelluloses and acids. The rate of decomposition of cellulose and the products formed vary with the type of organisms involved, the kind and amount of available nutrients, and the environmental conditions. When favorable conditions exist, cellulose tends to disappear rapidly. In well-decomposed material celluloses do not occur in very large amount.

Lignins are cell-wall constituents that are more resistant to decomposition by microorganisms than are plant carbohydrates. They are variable in chemical composition. The presence of lignin in plant materials is unfavorable for decomposition of cellulose. Waksman and Starkey (1931) stated that only certain Basidiomycetes and Actinomyces are capable of attacking lignin. *Tannins*, *waxes*, and *resins* appear to be resistant to decomposition by microorganisms. *Fats* may be decomposed to glycerol and fatty acids, and these are subsequently decomposed to carbon dioxide, water, and lower acids. Rehorst (1929: 246) regarded plant and animal fats and oils as relatively resistant to decomposition by microorganisms.

Proteins and their derivatives are highly important sources of nitrogen when decomposed by the soil organisms. Proteins are composed of amino acids which may be used by many soil organisms; upon decomposition of the amino acids the combined nitrogen may be assimilated or liberated as ammonia. Proteins are among the first substances to be acted on when plant materials are decomposed (Waksman, 1929: 195). Nonprotein nitrogenous substances, such as nucleic acids, alkaloids, purine bases, phosphatides, amines, and amides, may also occur in the organic material. The proteins in humus are partly of plant and partly

of microbial origin. *Mineral constituents* of organic matter, such as calcium, magnesium, potassium, phosphorus, and iron, play an important role in the nutrition of the microorganisms active in humus formation.

As decomposition of fresh plant material proceeds, the water-soluble compounds and the carbohydrates decrease rather rapidly. There is a much slower decrease, resulting in a relative accumulation, of fats and waxes, some hemicelluloses, and particularly the lignins and nitrogenous materials. There is also a relative increase in ash content if no losses result from leaching. Lunt (1933, 1935) found that during the first 7-8 weeks of weathering, after their fall in the autumn, tree leaves showed relatively little change in calcium content, but there was a distinct loss of phosphorus and potassium. Leaves of American beech showed the greatest loss of potassium and phosphorus; the loss of phosphorus was lowest in hickory. The loss of dry matter was greatest in red maple and flowering dogwood and least in American beech and oak.

Various constituents of plant materials have been regarded as the "mother substances" of humus. Evidently lignins are very important, but other constituents also play a role in humus formation either directly, in the form of hemicelluloses, proteins, and fats, or indirectly, through microbial synthesis (Waksman, 1938: 127). The processes which result in the formation of humus from fresh organic material are exceedingly complex. Decomposition and synthesis of compounds proceed side by side, with the result that chemical composition is continuously changing.

Decomposition of plant materials is influenced by many factors, such as the nature of the fresh material, the environmental conditions, and the microorganisms concerned. It would appear that climate exerts a primary control.

INFLUENCE OF THE NATURE OF THE FRESH PLANT MATERIAL ON DECOMPOSITION

Just as the chemical composition of fresh organic materials varies, so do the rate and type of decomposition. As a rule the debris of legumes decomposes more rapidly than that of non-legumes, and the litter of hardwood trees may decompose more

readily than that of conifers. Watson (1932) reported that samples of red maple leaves and Norway pine needles decomposed at different rates, as measured by carbon dioxide production. After decomposing for 2 months, red maple leaves had lost 27.5 per cent of their original dry weight, and Norway pine needles had lost only 14.0 per cent.

Differences in the decomposition of organic matter under adjacent stands of differing species composition have been observed repeatedly. In the first place, stand composition exerts an influence on the nature of the debris reaching the soil. Furthermore, the stand influences the microclimate, which in turn affects decomposition (Wittich, 1937). Even the lesser vegetation may be important in these respects (Ramann, 1898; Lang, 1926: 433; Hess, 1929; Wittich, 1937). There is a widespread opinion that the decomposition of organic debris under mixed stands is different, and ecologically more favorable, than under pure stands (Ramann, 1898; Wiedemann, 1925; Hesselman, 1927, 1937; Volger, 1933; Nömmik, 1938; Hwang, 1938). Numerous investigators have found that the decomposition of conifer debris is favored by the addition of hardwood leaves. Others have reported examples of more favorable conditions under pure stands of hardwoods than under pure stands of conifers (Wiedemann, 1925). Hesselman (1937) was of the opinion that there was a mutual effect between the vigor of a forest stand and the nature of the humus layers. With the development of unfavorable humus layer conditions vigor of the forest may decrease; on the other hand, a crowded stand lacking vigor can unfavorably influence the humus.

Although stand composition frequently may be influential in determining the rate and type of decomposition of organic debris, caution should be exercised in assigning it decisive importance in all situations (Hassenkamp, 1928; Romell and Heiberg, 1931; Aaltonen, 1932). Müller (1887: 161) stated that he knew of no vegetation whose litter always formed mor humus layers, but he recognized that some kinds of litter favor mor development to a greater extent than others. Factors other than the nature of the organic debris itself are frequently controlling. The influence of stand composition on humus layers may be expected to be greatest in regions where extreme climatic or geologic conditions do not rather rigidly predetermine humus layer types.

Debris from young trees or stands usually decomposes more readily than that from old trees (Albert, 1928; Weinkauff, 1928; Hesselman, 1937). Weinkauff found the decomposition of organic debris favorable in young stands on nearly all sites. With increasing age less favorable decomposition of organic matter appeared first on poor sites, later on medium-quality sites, and last, if at all, on the best sites. With increasing age the trees on all sites produced litter which was more resistant to decomposition. Hesselman (1937) found ammonia production, basic buffer content, and pH values higher in the humus in young stands than in old stands.

As a general rule, the richer in nutrient materials plant remains are, the more readily they decompose (Weinkauff, 1927). Nitrogen is a particularly important constituent, a relatively high content favoring rapid decomposition (Weinkauff, 1927; Watson, 1930; Waksman, 1932; Hwang, 1938). The presence of certain constituents, such as resins, appears to inhibit decomposition (Wiedemann, 1924). A high content of lignin in plant debris is also unfavorable; the lignins are not only very resistant to decomposition themselves, but also their presence inhibits the decomposition of cellulose (Waksman and Cordon, 1938). Grosskopf (1928) attributed the slow decomposition of spruce needles to their high content of lignin. High acidity and a low content of basic buffer substances in litter seem to be unfavorable to rapid decomposition (Lundegårdh, 1931: 366). Hwang (1938) reported a positive relationship between the pH value of litter and the pH of the decomposition residues.

INFLUENCE OF NUTRIENT CONDITIONS OF THE SOIL ON DECOMPOSITION

The chemical composition of the soil exerts a profound influence on the decomposition of organic matter and the nature of the humus layers (Wiedemann, 1924; Weinkauff, 1927; Dunnewald, 1930; Aaltonen, 1932; Wilde *et al.*, 1937). The nature of the soil, as has been indicated, influences the microorganic population and the composition of the organic debris which reaches the ground. Often site is probably more influential in favoring or inhibiting the development of more than is stand composition (Aaltonen, 1932).

It appears that, in general, the poorer a soil is in nutrients, the more difficult becomes maintenance of good humus layer conditions. Wiedemann (1924) and others have pointed out that decomposition of organic matter generally proceeds more satisfactorily in soils derived from basic rocks than in those derived from acidic rocks. Soils from basic rocks are usually richer in calcium and other nutrient elements and less acid.

Calcium is of outstanding importance in its influence on decomposition; its favorable effects are widely known. Grebe (1886: 160) pointed out that organic debris decomposes more rapidly in soils rich in lime than in those poor in lime. This observation also has been reported by Ebermayer (1890: 172), Jolyet (1916: 312), and Süchting (1926, 1929). In soils rich in calcium the tendency toward development of mor is not nearly as pronounced as in soils poor in this element (Hesselman, 1926; Krauss, 1926; Romell and Heiberg, 1931; Dengler, 1935: 282). Romell and Heiberg reported that the most extensive continuous areas with coarse mull in New York are found on soils rich in lime. Decomposition of organic matter is favored in soils which are under the influence of seepage water containing calcium (Hesselman, 1926; Hassenkamp, 1928). The beneficial effects of calcium may consist in part in neutralization of the decomposition products, improvement of the physical properties of the soil, and encouragement of an active bacterial population. Lang (1926: 339) has emphasized the importance of calcium in the decomposition of organic matter by bacteria. In northern regions an abundant supply of calcium in the soil or in seepage water may result locally in excellent humus conditions, which are ordinarily found only in more southern latitudes.

INFLUENCE OF AERATION CONDITIONS ON DECOMPOSITION

Under conditions of good aeration and favorable temperature and moisture, soil organic matter tends to decompose readily, whereas anaerobic conditions inhibit rapid decomposition. Waksman, Tenney, and Stevens (1928) showed that decomposition of oak leaves under aerobic conditions is more rapid than under anaerobic conditions. In the anaerobic preparations only the water-soluble constituents were attacked at first.

Lang (1926: 428) has suggested that a mixture of coniferous

needles and hardwood leaves may favor better aeration in the organic layers. It seems probable that the physical conditions in the organic debris layers exert considerable influence on their rate, as well as type, of decomposition. In fine-textured soils excessive amounts of moisture may prevail, and under these conditions decomposition of organic matter may be inhibited. Good aeration usually prevails in soils that are porous and not water-logged.

Seepage water containing dissolved oxygen appears to have a favorable influence on decomposition. This fact has been pointed out by various investigators, among whom may be mentioned Hesselman (1926) and Hassenkamp (1928).

INFLUENCE OF MOISTURE CONDITIONS ON DECOMPOSITION

Either deficient or excessive moisture is unfavorable for decomposition of organic matter (Müller, 1887; Hesselman, 1926; Süchting, 1929; Bornebusch, 1931). In Denmark, Müller found mor occurring more commonly in sandy soils than in those of finer texture; this same observation has been made in America by Romell and Heiberg (1931). Bornebusch (1931: 178) has shown that mor in Danish beech forests is in part a result of drying by wind. The unfavorable effects of sun and wind on humus conditions on south and west slopes in mountainous regions have been reported by Knickmann and Helbig (1925).

Wiedemann (1924: 14) recognized the unfavorable effects of deficient moisture supplies on decomposition and maintained that the average amount of moisture is not as important as the minimum amount during dry periods. An excessive content of rock in the soil may, in dry situations, favor drying out of the soil and hinder decomposition of the organic matter; on the other hand, in moist or wet mountain soils the influence of rocks may be beneficial, favoring more prompt decomposition (Grebe, 1886: 151). Ebermayer (1890: 171) stated that one reason for more active decomposition after thinnings is the increase in soil moisture to more favorable levels.

Excessive moisture is unfavorable to decomposition, since it leads to low temperatures and inhibits aeration. Lundegårdh (1931: 365) and Dengler (1935: 282) have pointed out that high moisture contents favor development of mor. Romell and

Heiberg (1931) emphasized the importance of ground-water conditions in determining humus layer types.

INFLUENCE OF TEMPERATURE CONDITIONS ON DECOMPOSITION

Temperature conditions influence decomposition of organic matter. This relationship is illustrated by the fact that decomposition proceeds rather slowly in northern latitudes, more rapidly in middle latitudes, and most rapidly in the tropics. Jenny (1930, 1931) and others have clearly shown that, as temperature increases, the content of organic matter in the soil decreases. On mountains decomposition is usually slower at high elevations than at low elevations. At and below freezing temperatures decomposition of organic matter ceases. Albert (1928) expressed the view that in litter decomposition, just as in soil formation, climate exerts a primary control.

Heavy thinnings which admit more solar radiation to the ground and increase soil temperatures have been advocated in regions of cool climate (Ebermayer, 1890: 171; Hassenkamp, 1928). However, the influence of thinnings on soil temperature appears to be quite variable. Wiedemann (1935) reported that, in general, heavy thinnings in beech and spruce stands in Germany had only slight effects on the humus layers of the soil. In heavily thinned beech stands the temperatures of surface soil during the summer were somewhat higher than in unthinned stands; the difference, however, was less than 1°C. A dense understory of tree reproduction or shrubs under a stand may exert an insulating effect, preventing solar radiation from reaching the soil after a thinning. Wiedemann (1935: 90) has pointed out that an understory in stands on some sites may lead to an accumulation of organic matter and an increase in acidity due to lowered temperature. Similar unfavorable effects have been reported as a result of underplanting with beech in pine stands in northwest Germany (Schneider, 1940).

Excessively high, as well as low, temperatures have an unfavorable influence on decomposition (Wiedemann, 1924: 15-17). The harmful effects of high temperatures are probably due in part to killing of soil microorganisms and in part to drying of the soil.

INFLUENCE OF SOIL ORGANISMS ON DECOMPOSITION

The influence of soil organisms on the nature of the humus layers formed in forests was emphasized by Müller (1887). In heath areas in Denmark he found that in certain favorable situations earthworms persisted for a long time after the forest disappeared and that they maintained a mull humus layer type. He ascribed great importance to earthworms in the development of mull in beech stands. In many soils a considerable part of the humus consists of animal excrement. Romell (1935) regarded decomposition in mor types of humus layer as primarily due to fungi and in mull types as primarily due to bacteria and animals. In mor considerable acidity appears to result from the activity of the fungi, but in mull the acidity produced is much less. The soil fauna was visualized by Romell as important in holding in check the acid-producing fungi; he referred to the soil animals as "guardians against biologically produced acidity."

CLASSIFICATION OF FOREST HUMUS LAYER TYPES

The term *humus layer* refers to the top portion of the soil, which owes its characteristic features to its content of humus. The humus may be incorporated or unincorporated in the mineral soil. Classification of forest humus layer types has been, and to some extent still is, in a state of confusion because of a number of circumstances, chief of which are: (1) loose and inconsistent nomenclature, (2) lack of information on the genesis and characteristics of the humus layer types being classified, and (3) varying bases of classification. The evolution of the terminology and classification of humus layer types has been discussed in detail by Romell and Heiberg (1931), Heiberg (1937), and Waksman (1938). These authors have brought out clearly the variations in nomenclature which exist. Thus the terms *raw humus*, *peat*, *upland peat*, *acid humus*, *torf*, *duff*, and *mor* all have been used in America to designate essentially the same type of humus layer. Similarly, *mild humus*, *leaf mold*, *neutral humus* and *mull* all have been used to designate another type of humus layer. Some standardization of nomenclature is clearly needed.

Müller (1878-1884, 1887), a Danish scientist, was probably one of the first to regard the humus layers in forests as natural biological units. His investigations led to the conclusion that the

nature of the humus layer is influenced by the kind of fresh organic debris, the locality or site, and the living organisms (plant and animal) present. Müller recognized two principal kinds of humus layer, *mull* (Danish, *Muld*) and *mor* (Danish, *Mor* or *Maar*), and characterized their biological and ecological properties. These two principal groups appear to be an adequate foundation for a classification meeting present needs. The distinctive features of mull are its diffuse lower boundary and its crumb-like or granular structure. In the mor group the humus layer is, as a rule, sharply delimited from the mineral soil material, unless this mineral material has been impregnated with organic matter that has been washed down. As a rule mulls are less acid than mors; furthermore, they usually show nitrification in nature, whereas mors generally do not, unless they have been activated by some disturbance (Romell, 1935). Fungi are the most important organisms in mor, but bacteria are more abundant in mull.

Both of these major humus types contain different forms, which may be regarded as subtypes or specific types. It is to be expected that transitional forms frequently occur. As an example of a typical form of mull, a type variously referred to as earthworm mull, worm mull, crumb mull, or coarse mull may be mentioned. This type is a porous, typically loose, friable mass with crumb structure and consists of an intimate mixture of humus and mineral soil. Coarse mull is usually inhabited by large earthworms and supports a rich vegetation of geophytes (Romell and Heiberg, 1931). Fibrous mor may be mentioned as a representative of the mor group. In this type the *F* layer is well developed; the entire humus layer is fibrous and more or less tough, but not compact. The lesser vegetation commonly includes members of the Ericaceae (Romell and Heiberg, 1931; Heiberg, 1937).

At this point it seems desirable to consider the strata or horizons which are frequently distinguished within the humus layer and the designations which are used in referring to them. In 1926 Hesselman recognized three strata, or layers, as follows:

1. Litter or *L* layer, the layer consisting of unaltered dead remains of plants and animals. Strictly speaking, litter is not humus or a part of the humus layer. Some workers have designated the litter layer as the A_{00} horizon.

2. *F* layer (*Förmultningsskiktet*), the layer consisting of partly decomposed organic matter. The structure of the plant debris is generally well enough preserved to permit identification of its source.

3. *H* layer (*Humusämneskiktet*), the layer consisting for the most part of well-decomposed, amorphous organic matter. Hesselman recognized an *H* layer in mull humus layer types as well as in mor types and stated that the *H* layer of mull was highly variable in thickness, always possessed a crumb structure, and contained more or less mineral soil material. Most American foresters prefer not to use the *H* layer designation in connection with mull humus layer types but rather to refer to the horizon in question as A_1 . Some writers have designated the combined *F* and *H* layers of mor types as the A_0 horizon.

At the Third International Congress of Soil Science, held at Oxford, England, in 1935, approval was given to a proposal for the nomenclature of forest humus layers by Bornebusch and Heiberg (1936). More recently the following revised nomenclature of forest humus layers has been proposed (Heiberg and Chandler, 1941):

Forest humus layers may be divided into two main groups: (I) mull and (II) mor.

I. *Mull*. A humus layer consisting of mixed organic and mineral matter. Transition to lower horizon not sharp. Types recognized in this group:

A. Coarse mull. Coarse crumb or granular structure. Many granules about $\frac{1}{8}$ in. (2.0–3.0 mm.) or larger. Organic matter thoroughly mixed with mineral soil (usually 5–20 per cent organic matter; exceptional cases even higher).

B. Medium mull. Medium crumb or granular structure. The larger granules about $\frac{1}{16}$ in. (about 2.0 mm.) or slightly smaller.

C. Fine mull. Granular structure, frequently having the appearance of fine black sawdust. Rich in organic matter (usually over 50 per cent, but sometimes as low as 30 per cent).

D. Firm mull. Dense, compact, and generally structureless, with a low content of organic matter (usually less than 5 per cent).

E. Twin mull. A complex type consisting of an upper horizon of fine mull or matted mor (see Group II below) underlain by an A_1 horizon having the characteristics of either medium or coarse mull.

II. *Mor*. A humus layer of unincorporated organic material, usually matted or compacted or both, distinctly delimited from the mineral soil

unless the latter has been blackened by the washing in of organic matter. Types recognized in this group:

A. Matted mor. *F* layer thin, sometimes practically absent. Organic matter of the *H* layer finely granular, as in granular mor or fine mull; when dry, virtually all of it can be shaken out from the dense root mat that holds it together.

B. Laminated mor. Thick, laminated *F* layer of matted leaves. *H* layer similar to matted mor.

C. Granular mor. *H* layer pronounced and of fine granular structure, lower part somewhat compacted. In dry condition, very easily broken into fine powder when crushed by hand. Distinguished from matted mor by the absence of a well-developed root mat.

D. Greasy mor. *F* layer usually weakly developed, commonly more or less fibrous. *H* layer thick and compacted, with a distinctly greasy feel when wet. Hard and brittle when dry.

E. Fibrous mor. *F* layer well developed; both *F* and *H* layers fibrous and more or less tough and felty, but not compact. Many plant remains may be visible in the *H* layer.

It should be recognized that the proposed terms apply only to forest humus layers of well-drained, upland soils, specifically to the northeastern United States. Additional humus layer types, as yet unrecognized, doubtless occur in other parts of America.

A proposal to use the term *forest floor* to designate all organic matter, inclusive of litter, on the mineral soil surface was adopted at the Third International Congress of Soil Science.

In general, mor is the most common type of humus layer in the colder parts of the temperate zone and at high elevations in mountains. Mor is commonly, but by no means exclusively, associated with coniferous forests on podzol soils. Even within the zone dominated by podzol soils, mull humus layer types may occur locally, for example, in situations where the soil is well supplied with calcium carbonate. Mull humus layer types are developed most commonly in regions lying south of the zone of podzol soils. Soils with mull generally, but not necessarily, support forests of hardwood trees. In regions where mull types are prevalent, mor may occur locally in situations which are excessively wet or dry and in areas where the soils are derived from parent materials poor in bases.

The humus layer types approved in 1935 and revised in 1941 are known to occur at various points in the northeastern United States, in the Middle Atlantic States, in the southern Appala-

chians, and on the Pacific coast. Bodman and Kittredge (1938) indicated the widespread occurrence of mull humus layer types in California forests.

Indicator plants may be of local value in suggesting the type of humus layer present (Müller, 1887; Romell and Heiberg, 1931; Lutz, 1932). However, extreme caution must be used until relationships between vegetation and humus layer types have been more carefully worked out than they are at present.

COMPOSITION OF FOREST SOIL HUMUS

It has been pointed out earlier and may be emphasized here that humus is not a specific substance but an exceedingly complex mixture of many compounds. Maiwald (1931: 116-117) has referred to it as a mixture of organic compounds of almost hopelessly complex nature. Composition varies continually, being influenced by the nature of the fresh plant and animal residues, the organisms active in the decomposition processes, the environmental conditions, and the stage of decomposition. Considered broadly, humus is made up of the following materials (Waksman *et al.*, 1928).

1. A number of residual constituents, such as celluloses, hemicelluloses, fats, and waxes of the plant and animal materials which are undergoing decomposition.

2. Constituents of the plant and animal materials which are more or less resistant to decomposition, for example, lignins, cutins, tannins, and resins.

3. Microbial cells synthesized in the process of decomposition of the organic matter reaching the soil.

4. Products arising from the decomposition of the original plant and animal materials and the synthesized microbial cell substance, for example, organic and inorganic acids, ammonia, nitrate, and mineral salts.

The acidity of humus material has long attracted attention; it results from the presence of both organic and inorganic acids. Numerous "humic acids" have been prepared by different investigators, and chemical formulas have been attached to them. These hypothetical acids have been considered in some detail by Waksman (1938). He concluded:

... one may, therefore, feel justified in abandoning without reservation the whole nomenclature of "humic acids," beginning with the "hu-

mins" and "ulmins" through the whole series of "humus," "hymatomelanic," "crenic," "apocrenic," and numerous other acids, and ending with the "fulvic acid" and "humal acids," the last additions to the list. These labels designate, not definite chemical compounds, but merely certain preparations which have been obtained by specific procedures.

Although exceptions occur, it may be stated as a general tendency that humus derived from debris of coniferous trees is more acid than that derived from debris of hardwoods. Alway, Kittredge, and Methley (1933) reported that the presence of basswood and maple leaves in mixture with the organic debris of jack, eastern white, and Norway pine stands in Minnesota resulted in a decrease of acidity in the organic layers. Data on the hydrogen-ion concentration of the organic layers under certain Minnesota stands are presented in Table 24.

TABLE 24. HYDROGEN-ION CONCENTRATION OF ORGANIC LAYERS OF SOIL IN MINNESOTA FORESTS

(pH values after Alway, Kittredge, and Methley, 1933)

Forest Type	Litter			F layer			H layer		
	Min.	Max.	Median	Min.	Max.	Median	Min.	Max.	Median
Jack pine	4.2	4.5	4.3	4.2	4.7	4.5	4.4	5.0	4.9
Norway pine	4.0	4.4	4.1	4.4	4.6	4.6	4.3	4.7	4.5
Eastern white pine	4.5	4.9	4.6	4.7	5.2	4.8	4.7	5.4	5.1
Maple-basswood	4.5	4.9	4.5	5.6	6.3	5.9	6.3	6.7	6.5

Lundegårdh (1931: 358) regarded mor as base-unsaturated and acid; mull was believed to tend toward base-saturation and to be only slightly acid. Romell and Heiberg (1931) found a rather steady decrease in pH value from coarse mull through the intermediate forms to fibrous mor. Hesselman (1926) reported that the material of the *F* layer of coniferous stands was generally less acid than that of the *H* layer. He also reported that the *H* layer usually had a lower content of basic buffer substances than the *F* layer. It appears that with increasing acidity of the humus layers, nitrogen content decreases (Hesselman, 1926; Němec, 1929). Němec (1926) reported that the amount of "humified" matter (determined by treatment with H_2O_2) in pine and hardwood forest soils in Czechoslovakia decreased with increasing acidity. Production of nitrate nitrogen also decreased with increasing acidity of the humus.

Waksman (1932, 1938) and others have pointed out that humus contains substantial amounts of organic nitrogenous compounds or protein-like complexes. Vitiņš (1924: 145) believed that the total nitrogen content of soil humus was usually about 5.0 per cent. Hesselman (1926) found the nitrogen content of forest humus to vary between the extreme limits of 0.6 and 4.3 per cent; usually the values were between 1.5 and 3.0 per cent. With rising pH values of the humus, nitrogen content usually increased. Aaltonen (1926) reported the content of total nitrogen in humus to be 1.8–3.0 per cent. In old-growth Douglas fir timber in Washington the forest-floor material contained an average nitrogen content of 0.87 per cent. Alway and Kittredge (1933) found the nitrogen content of the forest floor under aspen-paper birch stands in the Lake States to vary from 0.87 to 1.63 per cent, corresponding to 100–590 lb. of nitrogen per acre. In longleaf pine stands the litter contained 0.522 ± 0.027 per cent nitrogen, and the *F*-layer material contained 0.538 ± 0.022 per cent (Heyward and Barnette, 1936).

The carbon-nitrogen ratios of forest soil humus vary greatly. In most fresh plant residues the ratio is wide; that is, the proportion of carbon to nitrogen is high. As decomposition proceeds, the ratio narrows because of the liberation of carbon dioxide. In agricultural soils the carbon-nitrogen ratio of the organic matter is commonly about 10:1, but in forest soils it is usually much wider. The mean carbon-nitrogen ratios of the organic matter in various horizons of longleaf pine soils were approximately as follows: litter, 101:1; *F* layer, 47:1; *A*₁ horizon, 33:1 (Heyward and Barnette, 1936). Isaac and Hopkins (1937) found the carbon-nitrogen ratio in forest-floor material of Douglas fir stands to be about 57:1 and in the mineral soil about 24:1. With increasing depth below the soil surface the carbon-nitrogen ratios usually narrow. Morgan and Lunt (1932) reported carbon-nitrogen ratios in the organic matter of soils with mor to vary from 29:1 in the *F* layer to 20:1 in the *C* horizon; in soils with mull the ratios in the corresponding horizons were 33:1 and 15:1. Carbon-nitrogen ratios exceeding 20:1 are very common in forest soils. Even so, the carbon-nitrogen ratios of forest soil humus are much narrower than the ratios found in the fresh organic debris. Stebutt (1930: 155) reported that the carbon-nitrogen ratio of relatively fresh debris in forest stands may be as

wide as 250:1. It is evident that this ratio is so variable that the determination of nitrogen as a means of estimating carbon is useless. For a discussion of the carbon-nitrogen ratio in relation to soil classification the report of Anderson and Byers (1934) should be consulted.

In addition to its importance as a source of nitrogen, humus supplies considerable amounts of mineral elements needed by plants. Bujakowsky (1930) found that on a weight basis the unincorporated organic matter (*Auflagehumus*) contained substantially more potassium, calcium, and phosphorus than the underlying mineral soil. Samples of the organic layers and of the underlying mineral soil from pure stands of Norway spruce were subjected to electro-ultrafiltration (a combination of electrodialysis and ultrafiltration, water being used as the solvent) for a period of 7 hours to determine the amounts of potassium, calcium, and phosphorus which would go into solution. The organic layers contained 4 times as much soluble potassium, 8 times as much calcium, and 20 times as much phosphorus as the mineral soil.

In Table 25 are presented data on the composition of the organic layers of soil in various forest types in Minnesota. All stands were growing on Cass Lake fine sand.

TABLE 25. COMPOSITION OF THE ORGANIC LAYERS OF FOREST SOIL IN STANDS IN MINNESOTA

(After Alway, Kittredge, and Methley, 1933)

Organic Layer	Jack Pine	Norway Pine	Eastern White Pine	Sugar Maple-Basswood
<i>Volatile Matter, percentage of dry matter</i>				
Litter	95.52	96.60	93.15	88.14
F layer	83.02	88.15	81.67	82.37
H layer	47.49	57.68	56.64	51.07
<i>Calcium Oxide, percentage of dry weight</i>				
Litter	1.05	1.04	1.41	4.51
F layer	1.16	1.24	1.86	5.47
H layer	1.05	0.85	1.61	3.84
<i>Nitrogen, percentage of dry weight</i>				
Litter	0.87	0.69	0.94	1.17
F layer	1.35	1.15	1.41	1.93
H layer	0.98	1.07	1.41	1.82

TABLE 26. CHEMICAL PROPERTIES OF THE ORGANIC LAYERS OF THE SOIL IN DIFFERENT FOREST TYPES

(After Wilde *et al.*, 1937)

Type of Organic Debris	Reaction, pH	Base Exchange, m.e./100 gm. of soil	Total N, per cent	Available P, p.p.m.	Available K ₂ O, p.p.m.	Replace- able Ca, m.e./ 100 gm.
Jack pine, F layer	5.6 ± 0.1	13.3 ± 1.0	0.372 ± 0.034	24 ± 5.7	186 ± 26.7	7.5 ± 0.5
Norway pine, F layer	5.4 ± 0.2	29.3 ± 3.6	0.653 ± 0.060	38 ± 4.8	287 ± 14.2	12.9 ± 1.7
Eastern white pine, F layer	5.4 ± 0.2	47.1 ± 3.6	0.976 ± 0.110	60 ± 7.0	483 ± 32.9	28.2 ± 3.1
Eastern hemlock, F layer	4.8 ± 0.2	73.2 ± 8.6	1.490 ± 0.042	57 ± 3.4	432 ± 19.7	29.4 ± 4.7
Hardwood-hemlock, F layer	5.4 ± 0.4	72.5 ± 6.2	1.430 ± 0.070	148 ± 9.4	741 ± 29.0	44.1 ± 6.3
Hardwood, crumb mul	5.3 ± 0.1	27.6 ± 2.5	0.610 ± 0.050	37 ± 9.2	296 ± 19.0	16.5 ± 1.9

Table 26 contains the data of Wilde *et al.* (1937) on various chemical properties of the organic layers of the soil in different forest types of the Lake States.

In recent years the composition of forest soil humus has been investigated for certain groups of organic compounds. Illustrative data are presented in Tables 27 and 28. The ether- and alcohol-soluble complexes include fats, waxes, resins, sterols, higher alcohols, fatty acids, tannins, and pigments. Quantitatively these complexes constitute a relatively small proportion of humus substance. Their nature and abundance seem to depend on the type of forest vegetation, the degree of decomposition, and environmental conditions. Némec (1929) found the benzol-alcohol-soluble fraction was highest in humus from pine stands. He attributed this fact to the rather high resin content of pine litter. Némec reported an inverse relationship between nitrification and the content of benzol-alcohol-soluble complexes in humus samples incubated for 4 weeks.

TABLE 27. PROXIMATE COMPOSITION OF THE ORGANIC MATTER IN A COLTON SOIL PROFILE HAVING A MOR HUMUS LAYER (Data from an old eastern white pine-eastern hemlock stand near Keene, New Hampshire; values as percentages, on an ash-free basis.)

(After Watson, 1930)

Humus Constituent	Soil Horizon					
	<i>L</i>	<i>F</i>	<i>H</i>	<i>A</i> ₁	<i>A</i> ₂	<i>B</i> ₁
Ether-soluble fraction	6.7	4.2	3.9	4.6	3.4	1.0
Water-soluble fraction	5.4	5.1	4.1	4.7	4.4	2.0
Alcohol-soluble fraction	6.1	4.5	5.1	4.5	1.8	0.5
Hemicelluloses	13.6	11.7	10.6	9.9	7.0	3.8
Celluloses	13.7	8.2	5.2	2.6	1.0	1.1
Lignins	35.0	45.9	49.5	50.0	46.2	32.3
Protein	6.1	10.6	11.1	15.0	17.5	36.2

Carbohydrates and related compounds (cellulose, pentosans, galactans, mannans, polyuronides, starches, sugars, and glucosides) represent a major group of constituents in plants. In humus most of these substances are of less importance. Cellulose, a very abundant constituent in plants, may be present in very limited quantities in organic matter that has reached an advanced stage of decomposition. The *F* and *H* material of

TABLE 28. PROXIMATE COMPOSITION OF THE ORGANIC MATTER OF VARIOUS HUMUS LAYER MATERIALS (Values as percentage of dry weight)

Humus Constituent	Red Spruce, 75 Years Old, New Hampshire*			Northern Hardwood-Spruce, Maine†		Eastern White Pine Litter*	
	L	F	H	F	H	Fresh Material	After 2 Months' Decomposition
Ether-soluble fraction	5.15	2.64	2.36	4.41	2.99	6.1	4.2
Alcohol-soluble fraction	3.72	4.37	4.30			3.3	2.2
Water-soluble fraction	7.17	6.00	6.35	5.80	2.73	12.7	4.6
Hemicelluloses	14.0	14.3	12.6	15.28	12.39	14.5	12.6
Celluloses	16.3	10.6	4.17	9.44	2.56	15.8	10.7
Lignins	38.5	46.0	48.3	39.30	50.39	31.8	34.2
Protein	5.47	10.30	17.60	8.29	7.51	3.1	4.0
Ash				9.20	11.61		

* After Watson (1930).

† After Waksman *et al.* (1928).

mor humus layers, however, usually contains considerable cellulose. Most of the nitrogenous matter in humus is in organic forms, principally as microbial protein. The inorganic forms, such as ammonia and nitrate, constitute only about 2-3 per cent of the total nitrogen. It is definitely known that lignin-like complexes occur in humus in important amounts. The relative content of lignin usually increases as decomposition proceeds. Thus the lignin content in *F*-layer material is usually lower than in *H*-layer material.

AMOUNT OF ORGANIC MATTER IN FOREST SOILS

The organic matter of the soil represents an equilibrium or balance between the agencies supplying fresh organic debris and those leading to its decomposition. The balance is of course dynamic; the adjustment is probably closest in climax plant communities. Changes in natural conditions which upset the equilibrium between the two processes will result in either increase or decrease of soil humus.

Most of the organic matter which annually is added to the soil under forest stands is deposited on the surface. Data concerning the contribution made by roots which annually decay in the soil are lacking, but such contribution is believed to be small. In grasslands, by way of contrast, a large amount of organic matter is deposited annually in the soil body by the roots which

die. Desert soils have a low content of organic matter because of rapid decomposition of the small amount of plant debris which is added annually.

It is frequently stated that forest soils contain less organic matter than grassland soils and more than desert soils (Nikiforoff, 1938; Waksman, 1938: 127). Mosier and Gustafson (1917: 144-145) reported the average organic matter content in the first 6 $\frac{2}{3}$ in. of prairie soils in Illinois as 4.53 per cent, whereas the comparable value for forest soils was 1.93. However, there is some evidence that the alleged difference in amount of organic matter in grassland and forest soils may be fictitious. Dunne-wald (1930) investigated certain grassland and forest soils in the Bighorn Mountains in Wyoming and reported that, if the surface half-inch accumulation of needles and twigs in timbered areas is disregarded, the average difference in organic matter content of the soil beneath grass and timber is not great. Morgan and Lunt (1931, 1932) calculated the amount of organic matter in soils supporting the two types of vegetation and concluded that the differences are not great; in both kinds of soil the total organic matter content may vary from about 200,000 to 500,000 lb. per acre. They stated that the chief differences in the organic matter of grassland soils and forest soils relate not to amount but to character and vertical distribution.

Alway and Harmer (1927) investigated the weight of the forest floor (unincorporated organic debris) in nine virgin, or nearly virgin, forests on calcareous glacial drift in Minnesota. In the hardwood forests the weight of the dry organic matter averaged about 40,000 lb. per acre, varying from about 25,000 to 50,000 lb. In stands of conifers, dry organic matter averaged about 100,000 lb. per acre, varying from about 46,000 to 193,000 lb. In the hardwood stands the volatile matter averaged about 25,000 lb. per acre and in the conifer stands about 68,000 lb. Lunt (1932) presented data on the weight of organic debris, in pounds per acre, in various forest stands in New England. The maximum weight of *F*-layer material was 42,981 lb. per acre, based on samples from eastern white pine stands and Norway pine stands; the maximum for *H*-layer material was 263,547 lb. per acre for red spruce-hardwood stands in New Hampshire. Lunt suggested that the weight of the *F*-layer material, in terms of acre inches, is 15,000-20,000 lb.; of the *H*-layer material, about

40,000 lb. per acre inch. In 1933 Alway, Methley, and Younge reported on the amount of unincorporated organic matter in Northern Minnesota forest soils. Their results are presented in Table 29. The jack pine stand on the mainland was 59 years old and that on the island 100 years old. The Norway pine, eastern white pine, and maple-basswood forests were 60, 250, and 200 years old, respectively. The weight of the forest floor of aspen-paper birch stands in the Lake States varied from 8000 to 42,000 lb. per acre (Alway and Kittredge, 1933). In longleaf pine

TABLE 29. WEIGHT OF THE UNINCORPORATED ORGANIC MATTER IN MINNESOTA FORESTS (Values are pounds of oven-dry matter per acre)
(After Alway, Methley, and Younge, 1933)

Layer	Jack Pine		Norway Pine	Eastern White Pine	Maple-Basswood
	Mainland	Island			
Litter	3,111	5,376	5,875	3,476	2,037
F layer	19,570	11,215	12,640	24,970	8,200
H layer	19,330	16,860	19,089	40,910	49,550
Total	42,011	33,451	37,604	69,356	59,787

stands protected from fire for 10 years or longer the weight of the forest floor has been found to vary from about 20,000 to 55,000 pounds per acre (Heyward and Barnette, 1936). In the Douglas fir region on the Pacific coast the average weight of the forest floor has been reported to be about 64,000 lb. per acre (Isaac and Hopkins, 1937). Kittredge (1939) reported the dry weight of the forest floor in various chaparral types in California to vary from about 6600 to 46,300 lb. per acre. The data which have been cited indicate clearly that the amount of unincorporated organic matter in forests varies within extremely wide limits.

The amount of unincorporated organic matter, as distinguished from total organic matter, naturally is influenced by the type of humus layer. In mor types the amount of unincorporated organic matter is usually much greater than in mull types. The character of the mineral soil profile is another reason for variations in the weight of the forest floor. Alway and McMiller (1933) found greater accumulations of organic matter on mineral soils of fine texture than on those of coarse texture. Seasonal differences are also known to occur.

It has commonly been accepted as a fact that soils having a

mor type of humus layer contain more organic matter than those having a mull humus layer. The evidence at hand indicates that this is not the situation (Morgan and Lunt, 1931, 1932; Lunt, 1932; Romell, 1932). For example, Morgan and Lunt found the total organic matter per acre in two mor soils to vary from 271,081 to 554,450 lb.; in two mull soils the organic matter content varied from 222,627 to 515,231 lb. per acre. Calculations by Romell led to the conclusion that there was no consistent difference between crumb mull and pronounced forms of mor in the amount of organic matter stored in either the humus layer or the entire profile. It should be recognized, however, that under certain conditions soils with a mor type of humus layer may contain an amount of organic matter in excess of that in most mulls. For example, Romell (1932) reported the organic matter content of a soil in the Adirondacks having a greasy mor humus layer (40 cm. thick) as about 744,000 lb. per acre. This value represented the total organic matter content down to a depth of 1 meter.

Measurements of soil respiration under natural conditions in the field have been regarded as an index of the rapidity of decomposition of soil organic matter, and it has been claimed that soils with mull respire more strongly than those with mor. However, Romell (1932) took the position that determinations of soil respiration do not necessarily indicate anything about the rate of decomposition under balanced natural conditions. Even if soil respiration is accepted as a measure of rate of decomposition, there must be doubt that mull respire more strongly than mor. After critically reviewing available data on soil respiration and conducting investigations of his own, Romell (1932) found no indication that mull consistently respired more intensely than mor. In both humus layer types the average respiration amounted to about 0.2-0.7 gram of carbon dioxide per square meter per hour.

Mull humus layers in northeastern America commonly contain between about 10 and 20 per cent organic matter (determined by loss on ignition), based on dry weight; mor humus layers generally contain more than 80 per cent (Romell and Heiberg, 1931). In soils having a mull type of humus layer the percentage of organic matter decreases with increasing depth below the soil surface. In podzols with well-developed mor humus

layers the percentage of organic matter in the B_1 and B_2 horizons generally exceeds that found in the A_2 horizon.

The loss on ignition has commonly been used as an estimate of the content of organic matter in soil horizons containing relatively small proportions of inorganic matter, for example, the L , F , and H horizons of mor humus layer types and the L and F horizons of mull humus layer types. In horizons with a considerable proportion of mineral material it is better to calculate the amount of organic matter from the content of organic carbon by multiplying this percentage by the factor 1.724. This factor, which is widely used, is based on the assumption that the humus incorporated in the mineral soil contains 58 per cent carbon. Lunt (1931) has proposed using factors of 1.89 and 1.85 for calculating the organic matter content of litter material and F -layer material, respectively. Similar factors were derived by Heyward and Barnette (1936) in their work in longleaf pine soils.

ECOLOGICAL RELATIONS OF FOREST SOIL HUMUS

Humus is highly important from the standpoint of forest production. It functions in soil fertility, serving as a storehouse of nutrients necessary for plant growth. The physical, chemical, and biological properties of soil are improved by humus with the result that conditions for the growth of higher plants and microorganisms are made more favorable. Henry (1908: 55) suggested that the physical role of humus in soils may be even more important than its chemical role. Organic matter tends to make fine-textured soils more porous and binds together the particles in coarse-textured soils. Internal drainage and aeration of soils are facilitated by organic matter, and the field capacity is increased.

Mull types are generally regarded as more favorable for the growth of most forest trees, particularly hardwoods, than are mor types. However, it is possible, as Romell (1935) has suggested, that for some coniferous species mull may be too rich. It should not be assumed that forest production on mull types is always higher than on mor types. The type of humus layer is not the only factor which determines forest-site quality.

Development and maintenance of the highest possible level of forest soil fertility, as ends in themselves, are not the objective

of the silviculturist. His primary purpose is usually the production of those timber crops which most completely satisfy the particular objectives of management. Only in rare instances does the decision concerning the species to be grown and the cultural methods to be employed depend entirely on what is "best" for the soil. The fact that the coarse mull humus layer of a soil which formerly supported hardwoods may change when pure conifers, such as pine or spruce, are established on the land; should not be regarded as proof that such management is "bad." The changes which occur may actually render the soil better suited to the development of the pine or spruce; but, even if the changes are unfavorable, they must be accepted if the desired species are to be grown. It seems certain that humus layer conditions tend to be different under different forest types. Consequently, if the silviculturist decides to alter stand composition, he should anticipate that the humus layer, and possibly also the mineral soil profile, may exhibit change.

SOIL ORGANIC MATTER AND SITE QUALITY

In considering soil organic matter in relation to site quality, it is necessary at the outset to recognize that all soil organic matter is not the same. The two broad types of humus layer, mull and mor, are strikingly different ecologically.

Rapid decomposition of the organic debris reaching the soil is generally regarded as desirable (Albert and Penschuck, 1926; Weinkauff, 1927; Süchting, 1929; Krauss and Härtel, 1935). It has frequently been pointed out that tree growth in the tropics is rapid in spite of the fact that the mineral nutrient capital is low; prompt decomposition of the organic debris reaching the soil results in rapid turnover of the small nutrient capital. Ramann (1890: 101) believed that soil conditions are best when the organic debris reaching the ground decomposes fully within 1-2 years. Ijjász (1938: 229) stated that in humid climates the greater the annual variations are in the humus content in forest soils, the better the conditions for tree growth. Favorable effects of incorporated humus on site quality have been observed repeatedly, particularly in light-textured soils (von Falckenstein, 1912; Némec and Kvapil, 1927; Wiedemann, 1934: 48; Krammer, 1938).

Thick accumulations of unincorporated organic debris in the form of mor are frequently regarded as unfavorable from the standpoint of tree growth and soil conditions (Ramann, 1898; Hess, 1929: 30; Süchting, 1929; Süchting and Volkert, 1934: 296; Dengler, 1935: 282). Mor layers which increase appreciably in thickness throughout the life of a stand are usually considered especially undesirable. However, the thickness of the layers of organic matter is not as important as is the nature of the material;

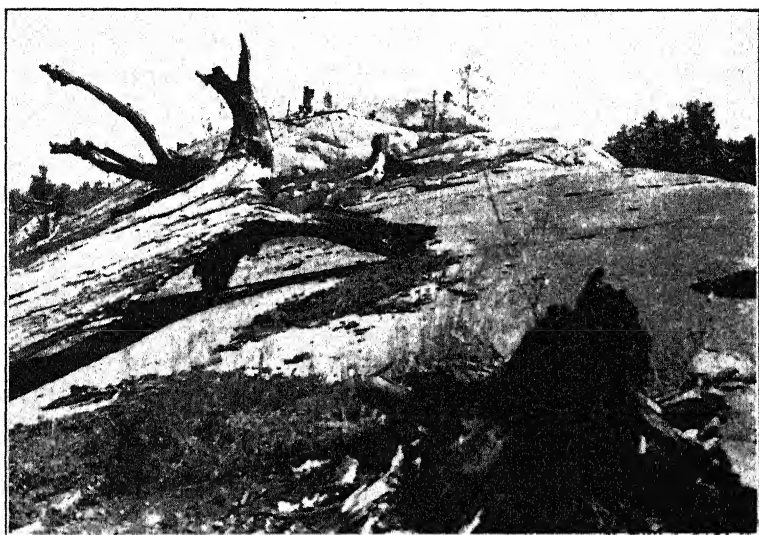


FIG. 23. Eastern white and Norway pine trees 30 in. or more in diameter once grew in organic debris on this rock outcrop in northern Michigan. Exposure and fires have destroyed the organic matter.

thick accumulations, at least in virgin forests, do not necessarily mean unfavorable conditions (Wiedemann, 1924: 40). In regions having a cool or cold climate, for example, at high elevations on mountains, trees are frequently found growing in purely organic soils which have developed on rocky situations where mineral soil is scanty or even lacking. Destruction of these organic soils by fire or exposure renders the areas nearly barren (Figure 23).

If organic debris builds up more rapidly than it decomposes, nutrient material is taken out of circulation. This fact has led some workers to designate the material of mor humus layers as

capital which bears no interest (Hassenkamp, 1928: 31). In addition, highly acid mor layers may result in the leaching of nutrients from the mineral soil and the development of undesirable biological conditions and physical properties. Soil structure may become poorer, pore volume may decrease, and aeration and water infiltration may be impeded. Considerable amounts of water may be withheld from the mineral soil by the unincorporated organic debris and lost by evaporation. The establishment of reproduction becomes difficult, since the young trees are rooted in the organic layers, which tend to dry out excessively during periods of drought.

MEASURES FOR IMPROVING HUMUS CONDITIONS

The humus layers of forest soils are more susceptible to influence and change by man than are the deeper-lying mineral soil horizons. In fact, they are the only horizons of the soil body which the forester can change rather quickly and without great expense. It should not be assumed, however, that it is always feasible, or even possible, to bring about such changes in the humus layers as may be desired. In regions where pronounced mor types are predetermined by extreme climatic or soil conditions the forester can scarcely expect to develop mull types. Changes in humus-layer conditions appear easiest to secure in situations where climate and mineral soil do not exert a controlling influence.

It is quite generally recognized that humus conditions may be influenced by altering stand composition. There is a widespread conviction that humus conditions under mixed stands are usually better than under pure stands. Frequently this appears to be a fact, but so many exceptions occur because of differences in the species involved that broad generalizations are unwise. European experience indicates that improvement of the soil is often obtained by introducing hardwood species into conifer stands of pine and spruce (Hesselman, 1926, 1927, 1937; Wiedemann, 1924, 1935: 90; Albert, 1924; Bornebusch, 1931; Tkatschenko, 1935; d'Aboville, 1936). These investigators have reported that in certain cases hardwoods mixed in conifer stands favor decomposition of the organic debris, reduce acidity, and stimulate nitrification. The beneficial effects of hardwood mixtures in spruce stands

in Sweden appear to persist even after the hardwoods have disappeared (Hesselman, 1927). In America Alway, Kittredge, and Methley (1933) have reported that the addition of maple and basswood leaves to pine litter increases calcium and nitrogen and reduces acidity.

In spite of the frequently beneficial results from hardwood mixtures in conifer stands, exceptions are known to occur; results are not always favorable (Hesselman, 1926; Chodzicki, 1930; Kmonitzek, 1930; Wiedemann, 1935: 90; Schneider, 1940). Schneider reported that beech was unsuited for underplanting in pine stands in northwest Germany. On sandy soils the leaves of beech were found to decompose slowly, they were poor in calcium, and the dense shade unfavorably affected soil temperature. Under these conditions organic matter accumulated, and the microflora and fauna of the soil decreased. Species with more deeply penetrating root systems than beech, and with a higher content of calcium, were found to have a favorable effect on the soil. Chodzicki (1930), working in a region on the southern border of the north German pine-beech-oak forests, reported that underplantings of beech in pine stands showed good development and had a favorable influence on the microorganic processes in the soil only on sites which originally supported pure beech or mixed pine-beech forest. Kmonitzek (1930) investigated the effect of underplanting spruce and beech in pine stands on sandy soils. Improvement was noted in certain physical and chemical properties of the soil, but soil moisture decreased as a result of the understory. This reduction in moisture was reflected in poor growth of the overwood.

By the use of thinnings the forester may sometimes appreciably change conditions in the humus layer of forest stands. More insolation and precipitation reach the ground as a result of thinnings, and in the openings created a more or less luxuriant minor vegetation may develop. Improvement of humus conditions by thinnings has been reported by various European investigators (Süchting, 1929; Bornebusch, 1931, 1933; Wiedemann, 1935). In America Adams (1935), Hansen (1937), and Spaeth and Diebold (1938) have indicated that thinnings result in somewhat higher temperatures in the humus layers. Hansen, who worked in jack pine stands in Minnesota, reported that humus-layer conditions in thinned stands appeared to be better

than in unthinned stands. Heavy thinnings have been found more effective in altering humus conditions than have light thinnings, and crown thinnings appear better than low thinnings. The changes frequently observed as a result of thinnings are in the direction of decreased acidity and increased nitrification and decomposition. The degree of change appears highly variable; in some cases it is substantial, in others negligible.

Clearcutting of stands has been found even more effective than thinnings in bringing about decomposition, lowered acidity, and nitrification in mor humus types. Wiedemann (1928) found that light thinnings in spruce had no significant effect on the acidity of humus layers, but full exposure by clearcutting has a pronounced effect. In this connection Wiedemann recognized that clearcutting may prove highly advantageous in some stands and regarded as unfortunate the opinion held by some foresters that clearcutting is always harmful. Of course, the belief that clearcutting is always acceptable, and never harmful to the soil, is equally unfortunate. It seems improbable that the better mull types of humus layer are ever improved as a result of clearcutting. In Sweden it has been found that exposure of the soil to insolation is all that is necessary to initiate nitrification in many mor humus types (Hesselman, 1917, 1927). Favorable effects on the soil resulting from clearcutting have also been reported by Němec and Kvapil (1927); Fraser (1929); Sachsse (1932: 247), and Duthie *et al.* (1937). Hesselman (1927) stated that the employment of long rotations seems to have an unfavorable influence on the soil in some instances. As a result proposals have been made in parts of both Germany and Sweden that the length of rotations be shortened.

Burning has been used to reduce mor layers which were thick and in unfavorable condition. Hesselman (1917) stated that in Sweden nitrification will not occur in certain mor types from simple exposure to the sun. It was found that nitrification was induced by burning. Nitrification was not regarded as indispensable for the establishment of regeneration but was considered desirable, since the process was then easier and the young trees developed more rapidly. Hesselman favored burning only when other measures failed to result in nitrification. The favorable effects of burning on the soil have also been considered by Wiedemann (1924), Süchting (1929), and Hesselman (1937).

The humus layers of a forest soil may be influenced by the method of slash disposal adopted. Various workers have advocated the scattering of slash on light-textured soils and on rather dry, stony soils (Albert, 1915; Dengler, 1928; Elpatievsky *et al.*, 1934; Wiedemann, 1934: 48; Fabricius, 1938). Benefits commonly claimed are increased organic matter in the soil, reduced moisture loss by evaporation, and better nutrient conditions. Improved color of the foliage and increased growth have been noted to result from additions of slash. Schulenburg (1936) used slash to mulch plantation stands of pine. This procedure has been tried in an experimental way on sandy soils in New York with promising results. Pearson and Marsh (1935) stated that in the Southwest a moderate cover of slash on bare soils which are either very fine-textured and compact or very coarse-textured has proven decidedly beneficial in aiding the establishment of ponderosa pine seedlings. Overgrazed clay flats, common on the Colorado Plateau, are typical examples of sites on which slash should be scattered for soil protection, as are the more or less bare cobblestone areas where the surface soil has been washed away. In the Engelmann spruce (*Picea engelmanni*) type deep layers of undecomposed organic debris are detrimental to both spruce and fir reproduction. On cold sites where such accumulations tend to occur the slash should be burned if this can be accomplished safely.

Mixing the organic debris with the mineral soil has been found beneficial in European forests (von Falckenstein, 1913; Wiedemann, 1924; Süchting, 1929; Tkatschenko, 1929). Decomposition and nitrification of the humus are favored by this measure. The principal objection to the treatment is its usually high cost. However, in areas where power machinery can be used it is possible that the cost would not be prohibitive. Practices such as liming and green manuring have found limited use in Europe but are not applied in America.

In 1935 Romell advanced a biological explanation of the activation effects frequently noted in mor humus layer types as a result of such forestry measures as intermediate or reproduction cuttings, burning, working the soil, liming, and green manuring. It was his belief that these measures tend, directly or indirectly, to weaken the fungus flora of mor humus layers and to stimulate development of a richer bacterial and animal life.

As a result of European experience it is known that forest soils are adversely affected by litter removal, especially when it is practiced annually (Schwappach, 1887; Henry, 1908: 91; Albert, 1924; Wiedemann, 1925, 1934, 1935; Krauss *et al.*, 1935). Litter removal, which has been a widespread practice in European forests, is encountered only in certain sections of America. One of these is the eastern shore of Maryland, particularly in the southernmost counties, where the litter from loblolly pine stands is commonly utilized for bedding purposes. Measurements made by Cope (1925: 69-70) in a 55-year-old stand of loblolly pine indicated that regular removal of litter over a period of years had resulted in lowering both height growth and volume. In the portion of the stand where needles were not removed, dominant trees averaged 78 ft. in height, and the total volume per acre was 24,800 board ft.; in the part where needles were removed, the dominants averaged 68 ft. in height, and the volume was only 18,600 board ft. More recently Jemison (1943) has reported a slight reduction of diameter growth in shortleaf pine as a result of annual litter removal over a period of 12 years from a mixed oak-pine forest near Asheville, North Carolina. Experience has clearly shown that litter removal from forest stands is incompatible with sound silvicultural management.

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Chapter 7

NATURE AND PROPERTIES OF SOIL COLLOIDS

The colloidal fraction of the soil is highly important from the standpoint of physical, chemical, and biological conditions. It is the most active part of the soil and in large measure determines soil character and productivity. In a given soil the colloidal material contains a very much larger proportion of the available plant nutrients than does the noncolloidal fraction. Because of its capacity to hold various chemical elements, colloidal clay is frequently compared to a storehouse in which the plant nutrients are conserved. The nature of the soil colloids is also of fundamental importance in studies of soil genesis and classification. Investigators are directing their attention more and more to this complex. It is essential that the student of forest soils acquire a working knowledge of soil colloids and the phenomena associated with them.

THE COLLOIDAL RANGE

Agreement concerning limits of the colloidal range is lacking, but the dispersed particles may be regarded as varying in size from the smallest in true suspensions to the largest in true solutions. They are too small to be seen with an ordinary microscope but may be viewed in the electron microscope. According to Ware (1936), the approximate limits commonly recognized by those interested in colloidal systems in general are from 0.2μ (0.0002 mm.) to $5 \text{ m}\mu$ (0.000005 mm.). The upper size limit of clay is usually set at 2μ (0.002 mm.), but many soil scientists regard 1μ (0.001 mm.) as about the maximum size for mineral colloidal particles. Consequently, all clay particles are not necessarily colloidal. Differences in definition of the upper size limit result from the fact that there is no sharp dividing line between colloidal and noncolloidal soil materials. In Table 30 are data on the average size of colloidal particles

derived from various soils. Because of the method of preparation the upper size limit of the material was about 0.3 μ .

TABLE 30. AVERAGE SIZE OF COLLOIDAL SOIL PARTICLES

(After Anderson and Mattson, 1926)

Kind of Colloid	Average Diameter of Colloidal Particles, millimicrons	Surface per Gram of Colloidal Particles, square meters	Number of Particles per Gram of Colloid, Divided by 10^{12}
Susquehanna	141	15.7	263
Norfolk	129	17.1	322
Sassafras	128	17.0	335
Huntington	111	20.4	505
Marshall	106	21.5	613
Sharkey	91	24.2	960
Fallon	102	21.3	680

Nelson and Hendricks (1943), using gas absorption measurements, estimated the specific surface of kaolinite as 15.5 sq. m. per gram of material. Values reported for other clay minerals were as follows: illite (hydrous mica), 97.1 sq. m. per gram; montmorillonite, 15.5; and halloysite, 43.2. The degree of aggregation of colloidal particles in soils may be regarded as intermediate between that of vacuum-dried materials as used for surface measurements and thoroughly dispersed materials.

In the past, soil colloid particles were regarded as more or less spherical, but the evidence now available points toward a preponderance of plate or rod shapes. Data concerning average diameters of particles should be considered with this fact in mind.

THE SURFACE AREA OF SOIL COLLOIDS

To a considerable extent the characteristics associated with colloidal substances are determined by their exceedingly large surface per unit of mass. Finely dispersed particles such as colloids have been referred to as being "nearly all surface." As a result of the tremendous surface areas involved, interfacial phenomena, such as solubility and adsorption, are prominent. Sand and silt fractions have relatively small surface per unit of mass and are rather inactive, both physically and chemically.

With increasingly finer subdivision of particles there is a very

large increase in the total surface area. If a cube of 1 cm. were divided into cubes of 5 μ (0.000005 mm.), the total area would amount to about 0.81 acre (Ware, 1936). One pound of colloidal particles (assumed to be spherical and to have a specific gravity of 2.7 and a diameter of 0.05 μ), if opened up and spread out as a flat surface, would cover an area of approximately 5 acres (Byers, Anderson, and Bradfield, 1938).

In considering soil colloids it is necessary to adopt the concept of an inner, as well as an outer, surface. This is particularly necessary in dealing with organic colloids and those inorganic colloids which belong to the montmorillonite group of clay minerals. The inner surface in inorganic colloids is primarily the space between the plates of the clay mineral crystals.

ORIGIN OF SOIL COLLOIDS

Soil colloids arise from two sources. The inorganic colloids result from physical and chemical alteration of minerals and rocks. Organic colloids originate from the plant and animal remains. The inorganic and organic colloids in the soil are intimately intermixed, at least in the upper soil layers, and together form the adsorption complex. The exact nature of the chemical changes which occur in minerals and rocks to produce inorganic colloids is not understood. In an effort to account for these changes, which are exceedingly complex, various theories have been advanced.

Physical disintegration and hydrolysis appear to be the primary agents responsible for the production of clay material (Byers, Alexander, and Holmes, 1935; Byers, Anderson, and Bradfield, 1938). The clay material may be regarded as consisting in part of residual products and in part of new products synthesized in the soil body. It is assumed that the intensity and extent of hydrolysis vary with environmental conditions, as does the nature of the new products which are synthesized.

The possible existence of the following fundamental compounds resulting from the hydrolysis of minerals has been postulated by Byers, Alexander, and Holmes (1935):

1. $3\text{H}_2\text{O} \cdot \text{Al}_2\text{O}_3 \cdot 6\text{SiO}_2$ (montmorillonitic acid)
2. $3\text{H}_2\text{O} \cdot \text{Al}_2\text{O}_3 \cdot 4\text{SiO}_2$ (pyrophyllitic acid),
3. $3\text{H}_2\text{O} \cdot \text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2$ (halloysitic acid)
4. $3\text{H}_2\text{O} \cdot \text{Al}_2\text{O}_3 \cdot \text{SiO}_2$ (allophanic acid)

In addition the existence of aluminum hydroxide and ferric hydroxide is recognized; these two compounds may be regarded as final hydrolytic products. The inorganic colloids are assumed to be salts of the acids listed above, with varying replacement of acid hydrogen by metals and of hydroxyl groups by acid ions, together with parent material of colloidal dimensions and unremoved decomposition products. Recent investigations have indicated that aluminum may replace silicon, and magnesium or iron may replace aluminum. Silicates of iron may be expected to show the same general behavior as those of aluminum, with the exception that they are hydrolyzed more readily than the aluminum compounds. It should be emphasized that no single inorganic colloid can be expected to consist wholly of the derivatives of any one of the acids mentioned. It also should be borne in mind that the colloidal system in any soil is not in a state of equilibrium but is undergoing change, the rate of which depends on the environmental conditions.

Mattson (1930, 1931, 1932a) considered the inorganic colloidal complex to result from mutual precipitation, at or near isoelectric conditions, of electropositive (basic) sols and electronegative (acidic) sols. Mattson advanced the view that the tendency in the weathering of soil minerals is always toward the formation of compounds possessing a maximum stability, both physically and chemically. This condition of maximum stability is represented by the isoelectric condition of the ampholytoids (amphoteric colloids), which in turn varies with the hydrogen-ion concentration. Thus under arid conditions, in the presence of abundant divalent bases, the colloids which originate will have a high silica-sesquioxide ratio.¹ Under cool, humid conditions the tendency is for the bases to be leached out and for high concentrations of hydrogen ions to develop. Under these conditions the colloidal complex that develops will have a somewhat lower silica-sesquioxide ratio. In regions with a warm, humid climate, just as in cool, humid regions, there is intense leaching of bases. However, in warm, humid climates there is rapid mineralization of the organic matter, and with the more intense weathering a

¹ In characterizing soil colloids the relative amounts of silica, iron oxide, and alumina have frequently been expressed in terms of chemically equivalent quantities. The ratio $(\text{SiO}_2/\text{Al}_2\text{O}_3 + \text{Fe}_2\text{O}_3)$ thus obtained has been called the silica-sesquioxide ratio.

small supply of bases is maintained and the reaction is more nearly neutral. Under these conditions of nearly neutral reaction the colloidal complex which is developed contains more sesquioxides and less silica; the result is a colloid with a very low silica-sesquioxide ratio. It may be pointed out that among other obstacles to acceptance of Mattson's theories on isoelectric weathering is the fact that the colloids in many American soils are not isoelectric; in most forest soils they are actually strongly electronegative.

The ideas just presented represent attempts to account for the formation of the secondary clay minerals in soils. It should be clearly recognized that the process is not well understood and that much more research on this phase of soil science is necessary.

OCCURRENCE OF COLLOIDS IN SOIL

The colloidal complex of the soil consists of organic and inorganic colloids which exist in part as a mechanical mixture and in part in chemical combination. The colloids coat the non-colloidal particles and also occur as independent components in the soil body. In forest soils the colloids usually exhibit a stronger tendency toward concentration in certain horizons than in agricultural soils. The organic colloids are generally concentrated in the uppermost layers and decrease with increasing depth. On the other hand, the inorganic colloids in most forest soils tend to be eluviated from the *A* horizons of the mineral soil and to become concentrated in the *B* horizons.

As a general rule, the noncolloidal fraction exceeds the colloids, but it is possible to have soils in which the colloidal fraction predominates. This latter condition exists in certain clay soils and in peats.

NATURE AND PROPERTIES OF SOIL COLLOIDS

All colloidal soil material does not behave the same. Inorganic colloids exhibit properties different from those of organic colloids. Furthermore, prominent differences are known to characterize various inorganic colloids, as well as organic colloids. Consequently, in addition to quantitative data on the amount of colloidal material in the soil, information concerning its nature and properties is also desirable.

INORGANIC COLLOIDS

Colloidal clay consists largely of the alteration products of silicate minerals in contrast to the coarse, noncolloidal soil fractions which consist for the most part of relatively unweathered mineral fragments. The proportion of undecomposed primary soil-forming minerals (resulting from mechanical disintegration) in the colloidal complex is highest in young soils and decreases with the advance toward maturity. For the most part the inorganic colloids exist as complex aluminum or iron silicates, together with varying proportions of bases.

In the past the inorganic soil colloids have frequently been held to be predominantly or exclusively amorphous (Schucht, 1930: 88). However, in 1928 Ross found by means of X-ray investigation that most of the clay-forming minerals are definitely crystalline with fairly well-defined chemical and physical properties. Subsequently Hendricks and Fry (1930) examined with X-rays colloidal material (maximum size of particles being 0.001 mm.) from a wide variety of soils and also found it crystalline. Further evidence on the crystallinity of soil colloids was advanced by Kelley, Dore, and Brown (1931), Marshall (1935), and Kelley (1936).

Additional proof of the crystalline nature of colloidal clay has more recently been obtained through use of the electron microscope by Shaw (1942) and others. Two electron micrographs are shown in Figure 24. In the montmorillonite, on the left, the material is crystalline and plate-shaped. The micrograph of kaolinite at the right shows that crystals of this mineral are flat and hexagonal in outline. Much greater regularity is exhibited by the kaolinite crystals than by the montmorillonite crystals. It is Shaw's opinion that montmorillonite, when thoroughly dispersed in water, exists as unit crystal plates (approximately 1 μ in thickness); a particle-size distribution curve for a given montmorillonite sample represents only the degree of dispersion.

Most authorities now recognize three main groups of clay minerals: kaolinite, montmorillonite, and hydrous mica. In view of their importance a brief discussion of the nature of the minerals in each of these groups is desirable.

Kaolinite is a hydrous aluminosilicate having the formula $(\text{OH})_4\text{Al}_2\text{Si}_2\text{O}_5$. The principles of the atomic structure of min-

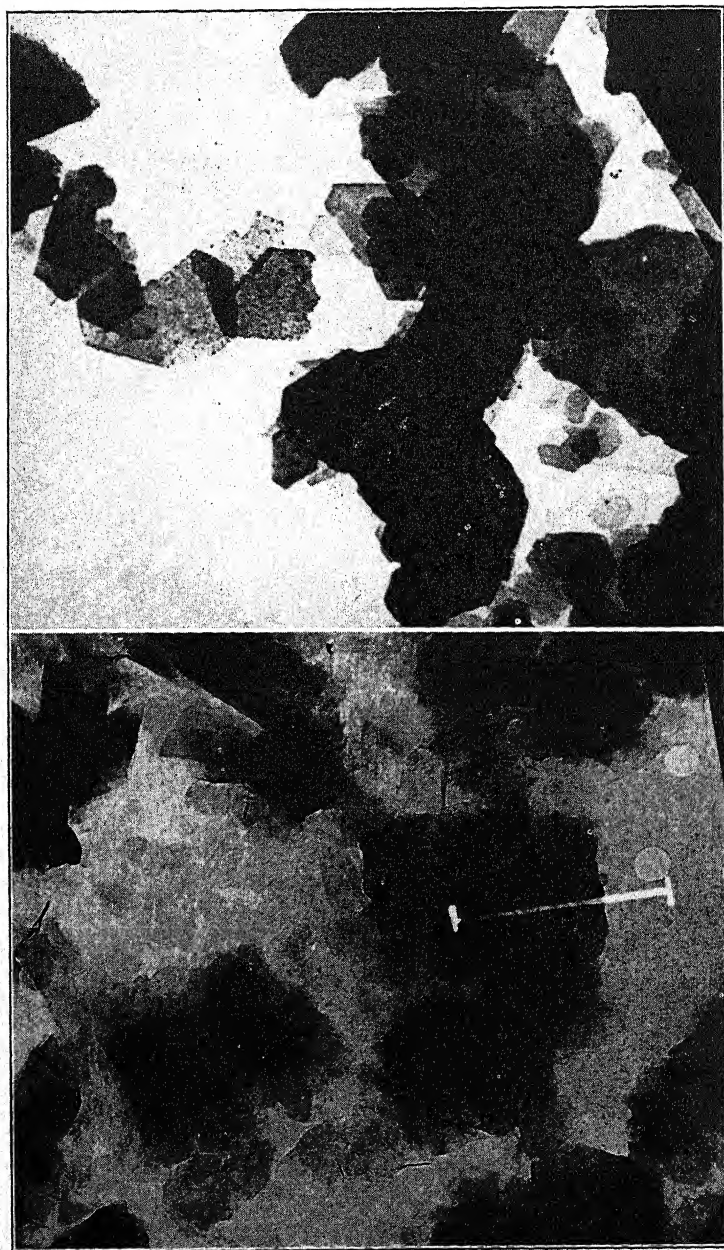


FIG. 24. Electron micrographs showing montmorillonite on left (2-1 μ fraction) and kaolinite on right (5-2 μ fraction).
Photographs by B. T. Shaw.

erals presented in Chapter 2 are applicable here to the structure of the secondary clay minerals. Thus the kaolinite molecule has a sheet of silicon-oxygen tetrahedra linked together by common oxygen ions at the corners (Figure 1C). There are four silicon ions for every ten oxygen ions. The tetrahedra being pointed upwards, four of these oxygen ions are apical and are shared with a sheet of linked alumina octahedra above. This second layer (gibbsite sheet) contains four aluminum ions, linked below to two OH ions in the same plane as the four apical oxygen ions of the silicon tetrahedral sheet, and linked above to six OH ions. A schematic presentation of the crystal structure of kaolinite is given in Figure 25. It should be noted that the kaolinite crystal con-

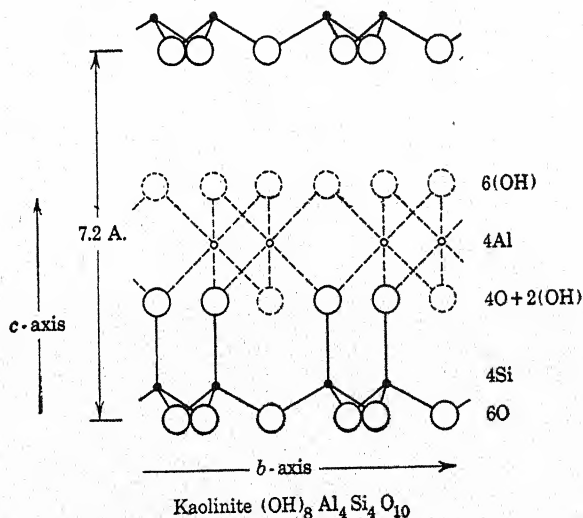


FIG. 25. Schematic representation of the crystal structure of kaolinite.

sists of a silica sheet and a gibbsite sheet, held together by linkage through common oxygen ions. Because of this structure the mineral is considered to be of the 1:1 type; it is referred to as having a fixed crystal lattice. Attention is also directed to the fact that there is no substitution of cations in the crystal; that is, aluminum does not substitute for silicon in the tetrahedral position, and magnesium or iron is not substituted for aluminum in the octahedral position. The kaolinite crystal is a balanced

structure and hence has no residual negative charge. In the absence of substitutions there is no distortion of the lattice, and the crystals can grow in any direction; unlike the montmorillonite clay minerals, kaolinite does not readily break down to thin sheets. Kaolinite is considered to have a fixed lattice and exhibits little shrinking and swelling with moisture changes, apparently because any two unit cells of kaolinite are held together by hydrogen bonding between the OH groups of one plate and the adjacent oxygen ions of the other. The characteristics mentioned for kaolinite should be borne in mind, for quite different conditions will be shown to exist in minerals of the montmorillonite group.

The minerals dickite and nacrite may be mentioned as other representatives of the kaolinite group of clays. These two minerals differ from kaolinite chiefly in that the position of the silica and the gibbsite sheet in relation to each other has undergone a shift or rotation.

Kaolinite was formerly considered to be the characteristic mineral of the clay fraction in soils. It is now known that the significance originally attached to this mineral was exaggerated; kaolinite is only one of a number of important clay minerals. It occurs in mature soils of both humid-temperate and humid-tropical regions.

Montmorillonite and the related clay minerals have received considerable attention in recent years. They exhibit more pronounced colloidal properties than kaolinite, have a higher exchange capacity, swell to a marked degree on wetting, and, in general, constitute a very active part of the soil body.

It is believed that molecules of montmorillonite are built up of two silicon-oxygen tetrahedral sheets and one gibbsite sheet. The gibbsite sheet lies between the two layers of silicon tetrahedra. Minerals having this structure are said to have a 2:1 lattice instead of the 1:1 lattice characteristic of kaolinite. The montmorillonite clay minerals are also referred to as having an expanding crystal lattice. The unit crystal plates, being less firmly linked than in the kaolinitic clays, which have a fixed lattice, expand and contract with changes in the amount of water present. In kaolinite one side of the gibbsite sheet is not in contact with a silica sheet and therefore presents exposed hydroxyl groups to the dispersion medium. Montmorillonite, on the other

hand, is believed to display few, if any, hydroxyl groups to the dispersion medium except on plate edges. The significance of this fact from the standpoint of colloidal properties of the two minerals will be considered later.

The exact structure of montmorillonite is still uncertain, but most authorities appear to favor that proposed by Hofmann, Endell, and Wilm (1933).² This structure is shown schematically in Figure 26. The vertices of the silica tetrahedra in both

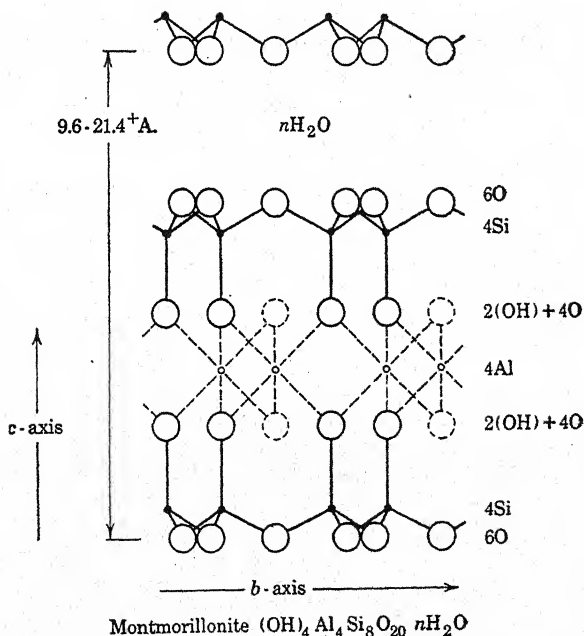


FIG. 26. Schematic representation of the crystal structure of montmorillonite.

silica sheets point inward toward the gibbsite sheet. Ideally only oxygen ions (no hydroxyl groups) are considered to be exposed to the dispersion medium. Actually in crystal fragments

2 Edelman and Favejee (1940) have proposed a structure for montmorillonite which differs from that of Hofmann *et al.* by having every other silicon-oxygen tetrahedron inverted in both silica sheets. In this structure OH groups are exposed to the dispersion medium. The structure proposed by Edelman and Favejee aids in explaining how cations may be held on the basal planes of montmorillonite crystals.

some of the OH groups of the gibbsite sheet are exposed along the flake edges.

The structure shown in Figure 26 indicates a balanced, neutral crystal. In reality there appears to be a lack of charge balance within the lattice structure of montmorillonite. This lack of charge balance, as explained in the section on cation exchange, is presumed to result from the substitution of certain cations in the tetrahedral and octahedral positions. Hendricks (1942) has suggested that the substitutions within the crystal lattice of the 2:1 types of clay minerals cause distortion and are unfavorable to the development of large crystals.

Montmorillonite clays seem to be most prominent in soils developed in regions of relatively low rainfall. They are known to occur in soils of the prairies, plains, and deserts.

The *hydrous micas* represent a group of clay minerals that are not well characterized. Some writers refer to them as illite. The illite structure is similar to that of montmorillonite except that about 15 per cent of the silicon positions in the tetrahedral sheets are occupied by aluminum, and the resulting excess negative charges are satisfied by potassium ions that are supposedly located between the silica sheets of two successive units (Grim, 1942). The potassium ions seem to hold the units together so that they do not expand in the presence of water. Comparison of the structure of hydrous mica with that of muscovite reveals considerable similarity. Some workers consider the hydrous micas as having been formed directly by a slight alteration of muscovite. This group of clay minerals may be prominent in the forest soils of regions having a cool, humid climate and are also known in soils of the prairies, plains, and deserts.

Within the three general groups of clays there are many different minerals. The individual members of a given group are differentiated largely on the basis of the kind and degree of cation substitution in the crystal lattice or by differences in crystallographic forms.

In the foregoing discussion consideration has been directed to the aluminosilicate clay minerals which are so important in determining the physical and chemical properties of soils. There is another group of inorganic soil materials which fall within the size range of clays but which differ considerably in their properties. These are the iron and aluminum oxides and their hydrated

forms. The composition and structure of some of these materials were considered in Chapter 2. They are present in considerable quantity in the *B* horizons of podzol and podzolic soils, most commonly as coatings on the larger mineral particles. In soils of subtropical and tropical regions where weathering has been intense they are particularly abundant, there comprising a large proportion of the soil material. Soils thus constituted are not highly plastic, do not shrink and swell appreciably with changes in moisture content, and have a low cation exchange capacity.

Although detailed information on the geographical distribution of clay minerals is lacking, it appears that neither climate nor parent material is an absolute control. In a given soil the inorganic colloidal material is usually composed of not one clay mineral but several.

CATION EXCHANGE

Undoubtedly the most important characteristic of colloidal clay is its property of cation exchange.³ Work on cation exchange was initiated in the middle of the nineteenth century with the discovery of Thompson (1850) and Way (1850) that soil was capable of removing ammonia from solutions. Two important facts were early established: (1) that there was a general chemical equivalence between adsorption and displacement of ions, and (2) that the exchange phenomena were associated with the colloidal material. For a historical review of the early development of the subject the papers of Prescott (1916) and Gedroiz (1918) should be consulted.

By way of introduction to the subject of cation exchange certain concepts concerning the situation at the interface of a colloidal particle and the dispersion medium need consideration. When colloidal clay particles are suspended in water, some dissociation of ions occurs. The anionic groups are fixed in the crystal lattice of the clay particle, but the dissociated cations occupy positions at varying distances from the corresponding surface anions. The situation which results is commonly portrayed as an electrical double layer. The inner layer is represented by the negatively charged anions in the crystal surface, and the

³ Many writers refer to this property as base exchange. The term *cation exchange* seems preferable because hydrogen is involved, as well as the metallic cations.

outer layer by the positively charged cations, which exist as a swarm or cloud around the clay particle. Figure 27 is a diagrammatic representation of a clay particle with an ionic double layer.

Cation exchange is, in short, a process whereby cations of the clay crystal are exchanged for cations in the soil solution or dispersion medium. In the process a cation from the soil solution is adsorbed, and another cation passes into the soil solution.

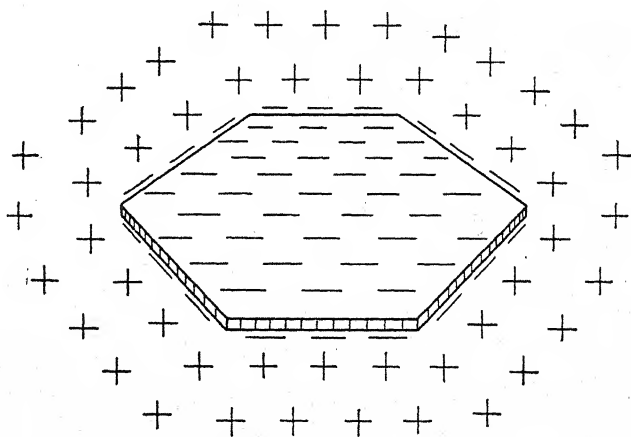
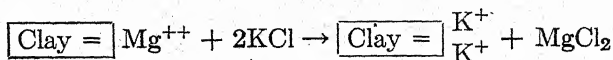


FIG. 27. Diagrammatic representation of a negatively charged colloidal particle. Cations are accumulated on the surface of this "colloidal anion."

Replacement may occur by virtue of either (1) the fact that the entering ion is adsorbed more strongly, that is, has a high replacement capacity, or (2) mass action effect. The process in simple form may be represented diagrammatically as follows:



It should be remembered that these reactions are reversible. However, by removing the liquid in equilibrium with the colloidal particles and replacing it with fresh solution, the particles may be saturated with one ion, and the reaction thereby brought to completion. Thus, if a given amount of soil is treated with a small amount of $\text{Ba}(\text{CH}_3\text{COO})_2$, an equilibrium is established

between the liquid and the colloidal particles. The colloidal complex contains some Ba ions but is not saturated with them. If, however, the soil is slowly leached over a period of several hours with a solution of $\text{Ba}(\text{CH}_3\text{COO})_2$, it is possible to replace all exchangeable cations with barium ions.

Some of the common cations may be arranged in accordance with their usual strength of adsorption, or replacing capacity, as follows: $\text{H} > \text{Ca} > \text{Mg} > \text{K} > \text{Na}$. The greater the energy of adsorption of an ion, the less readily it will be washed out of the soil; it is significant that of the bases mentioned, calcium is generally most abundant and sodium least abundant in soils of humid regions. Giesecking and Jenny (1936) arranged certain monovalent cations in the order of their strength of adsorption: $\text{H} > \text{Cs} > \text{Rb} > \text{NH}_4 > \text{K} > \text{Na} > \text{Li}$. These authors pointed out that divalent ions, unlike monovalent ions, do not produce a consistent adsorption series. For example, they found that the adsorption series in NH_4 clay is $\text{Mg} \begin{smallmatrix} \leq \\ \geq \end{smallmatrix} \text{Ca} < \text{Sr} < \text{Ba}$, whereas in H clay the series is $\text{Mg} < \text{Ba} < \text{Ca} < \text{Sr}$. In regard to release the arrangement of the foregoing ions in the first two series is opposite to the arrangement based on adsorption; that is, the most easily introduced ion is the most difficult to replace. It appears that both adsorption and release of ions are influenced by valence. As a general rule, the divalent ions have higher replacement power and are more tightly held than monovalent ions. Exchangeability series such as have been given do not hold exactly for all clay minerals. Probably each clay mineral has its own series.

It has commonly been stated that the replaceability of cations is a function of their hydration. Thus highly hydrated ions have been regarded as being held loosely and hence being easy to replace, whereas ions in a low state of hydration have been thought to be held more firmly and consequently to be more difficult to replace. Grim (1942) has recently expressed the view, however, that cation exchangeability cannot be satisfactorily explained on the basis of cation hydration. The reason for this position, which appears justified, will be considered in the section dealing with flocculation of colloidal clays.

Cation-exchange reactions are generally very rapid and, as earlier noted, reversible. The rapidity with which the reactions occur is to be expected, since they are surface reactions. Pall-

mann (1939) has pointed out that the rate of exchange is more rapid in systems where ion exchange takes place only on the outer surface, as in kaolinite, than in systems having porous particles, as in humus.

The total capacity of soil colloids for holding cations is referred to as the cation-exchange capacity. This capacity is usually expressed in terms of milligram equivalents⁴ per 100 grams of soil. When the exchange capacity is satisfied by metallic cations, the soil is considered to be base-saturated. This condition seldom exists in forest soils of humid regions because the bases are continually being replaced by hydrogen ions, with the result that the soil tends to be base-unsaturated. If the replacement is complete, the soil becomes hydrogen-saturated. The content of exchangeable hydrogen plus the content of exchangeable metallic cations equals the total cation-exchange capacity. The data in Table 31 illustrate the relationship of the three values. In humid regions base unsaturation may even exist in soils which contain

TABLE 31. CATION-EXCHANGE CAPACITY, EXCHANGEABLE HYDROGEN, AND EXCHANGEABLE METALLIC CATIONS OF CERTAIN FOREST SOILS

Soil Type and Horizons	Cation Exchange Capacity, m.e. per 100 gm. of soil	Exchangeable Hydrogen, m.e. per 100 gm. of soil	Exchangeable Metallic Cations, m.e. per 100 gm. of soil
Lakewood sand*			
A ₁	3.00	2.00	1.00
A ₂	0.77	0.55	0.22
B ₁	2.74	2.60	0.14
B ₂	2.74	2.55	0.19
C	0.64	0.50	0.14
Hermon loamy sand†			
L	67.93	57.93	10.00
F	100.00	96.83	3.17
H	105.41	104.37	1.04
A	16.51	14.05	2.46
B ₁	7.51	7.35	0.16
B ₂	3.62	3.27	0.35
C ₁	1.26	1.09	0.17
C ₂	0.90	0.51	0.39

* From southern New Jersey (Joffe and Watson, 1933).

† From southern New Hampshire (Lutz *et al.*, 1937).

⁴ One milligram equivalent (abbreviated m.e.) signifies the equivalent weight, that is, the atomic weight divided by the valence, taken in milligrams. For example, the atomic weight of calcium is 40.07 and its valence 2; therefore 20.035 mg. of calcium is 1 m.e. In the same way 1 mg. of hydrogen is 1 m.e.

free calcium carbonate in the form of relatively hard, coarse fragments of limestone. Although base saturation prevails immediately adjacent to the limestone fragments, the soil body as a whole is unsaturated because of loss of exchangeable bases by leaching (Robinson, 1936: 112).

In the soils of humid regions the most abundant exchangeable cations are: first, H and Ca, second, Mg, third, K and Na; in well-drained soils of arid regions the most abundant are: first, Ca and Mg, second, Na and K, third, H (Lyon and Buckman, 1943: 72). Schairer (1937) found that ammonia (NH_4) represented, on the average, about 30 per cent of the total exchangeable bases in the organic layers of forest soils examined in Germany.

The cation-exchange capacity of soil colloids varies greatly. As a general rule, inorganic colloids have a lower capacity than organic colloids, when compared on the basis of equal weights. Robinson (1936: 121) cited values of 16.4–110.2 m.e. per 100 grams for inorganic colloids and 250–450 for organic colloids. Powers (1932) found that *F*-layer material from stands of redwood (*Sequoia sempervirens*), Douglas fir, and ponderosa pine had a cation-exchange capacity of about 100 m.e. per 100 grams. The *F*-layer material investigated contained about 65 per cent organic matter. Samples from ponderosa pine stands on the east slope of the Cascades showed a higher cation-exchange capacity than samples from Douglas fir stands in the humid coast region. Bayer (1940: 97) indicated that colloidal humus has an exchange capacity of about 400 m.e. per 100 grams, in contrast to 60–100 m.e. per 100 grams for colloidal clays. As will be explained later, differences between exchange capacities of organic and inorganic colloidal materials are less when volume, rather than weight, is used as the basis for comparison.

The cause of the exchange capacity possessed by inorganic clay colloids is still not thoroughly understood. At present the concept most generally held seems to be that in minerals with a 2:1 type of lattice the exchange capacity results from the presence of a high negative charge on the clay crystal. This negative charge, which is necessary to hold exchangeable cations, arises from the substitution in varying degree of relatively low-valence cations for relatively high-valence cations within the octahedral and tetrahedral groups in the crystal. When Al^{+++} substitutes for Si^{++++} in the tetrahedral silicon sheet, an excess

negative charge results, since silicon has four positive charges and aluminum only three. This negative charge is compensated, or balanced, by cations which are taken up; adsorbed cations are for the most part exchangeable. In the same manner an excess negative charge results when Mg^{++} substitutes for Al^{+++} in the octahedral aluminum sheet. Aluminum in octahedral coordination may also be substituted by Fe^{++} or Fe^{+++} . Another substitution which occurs occasionally is that of Li^{+} for Mg^{++} (Ross and Hendricks, 1942). This substitution is possible because the crystal ionic radii of the two ions are about the same. Calcium, potassium, and sodium do not occur in the crystal lattices of the secondary clay minerals. Their size is too great to permit inclusion within the tetrahedral and octahedral groups.

These concepts as to the cause of exchange capacity in colloidal clay apply rather well to three-layered (2:1) clay minerals. Difficulties are encountered, however, when an attempt is made to apply them to kaolinite and other two-layered (1:1) clay minerals, because kaolinite is a balanced unit without any substitutions. Kelley and Jenny (1936), Shaw (1942), and others have attributed the exchange capacity of kaolinite to the exposed hydrogen ions of the gibbsite sheet (aluminum octahedra). It is thought that the hydrogen of the OH groups can exchange with other cations, thus accounting for the exchange capacity which samples of kaolinite usually show. It is known that the normally low exchange capacity of kaolinite may be greatly increased if the mineral is ground to extreme fineness. Presumably this increase of exchange capacity is due to the exposure of more and more OH groups as the mineral crystals are split up. It is also possible that broken oxygen bonds may play a role in cation exchange.

It thus appears that, when there is a substitution of relatively low-valence cations for relatively high-valence cations in the lattice of a colloidal clay crystal, cation-exchange capacity results. It must not be assumed, however, that clay minerals lacking these substitutions, such as kaolinite, also lack exchange capacity.

ANION EXCHANGE

In contrast to the substantial fund of knowledge concerning cation exchange, relatively little information is available on

anion exchange. A few brief remarks, however, can be made.

Adsorption of anions by colloidal clay is of particular interest from the standpoint of phosphorus relations in soils. The fact that phosphorus is commonly fixed in the soil in forms not readily available is well known. The explanation, although still somewhat obscure, seems to be that this fixation results from chemical precipitation and from adsorption by soil colloids. In connection with such adsorption it appears that PO_4 ions may be exchanged with OH and perhaps other ions in exchangeable positions on the surface of colloidal clay particles. Thus the capacity of kaolinite to fix phosphorus is attributed to replacement of exposed OH ions by PO_4 ions. Although phosphate ions may be adsorbed over a rather wide range of soil acidity, the greatest amount of adsorption appears to occur at low pH values.

SWELLING OF COLLOIDAL CLAYS

Changes in the water content of colloidal clays are accompanied by variations in volume. The amount of change varies with the nature of the clay, being substantial in some clays and slight in others. The addition of water results in swelling; and conversely, its removal leads to shrinkage. Water adsorbed by colloidal clay is compressed as a result of the attractive forces in the surfaces of the mineral crystals. This accounts for the fact that, although clay expands on being wetted, the total volume of the system (clay plus water) contracts. In other words, 1 cc. of solid colloidal clay added to 1 cc. of water results in a total volume of less than 2 cc. Although this volume contraction is thought to result chiefly from the change in volume of the liquid, it is probable that in some colloidal materials there is also an appreciable change in the volume of the solid (Katz, 1933).

The degree of swelling exhibited by colloidal clays bears a relationship to the kind of clay mineral involved. Thus the montmorillonite group of clays show much greater swelling than do members of the kaolinite group. Baver (1940) has presented data indicating that hydrogen-saturated montmorillonite and beidellite (both 2:1-type clays) show swelling of 2.2 and 0.81 cc. per gram, respectively. Halloysite (a 1:1-type clay) shows swelling of only 0.05 cc. per gram. The pronounced swelling of the montmorillonitic clays is usually explained by the expanding

lattice these minerals possess. Water molecules enter between the individual crystal plates and push them apart; several molecular layers of water may be built up within the interplate spaces. Thus in clay minerals of the montmorillonite type water is adsorbed on the interplate surfaces within the crystals as well as on the outer surfaces. The kaolinite type of clay minerals, on the other hand, have a fixed lattice which does not permit entry of water into the crystals, with the result that adsorption occurs only on the outer surfaces. In the montmorillonite type of clays swelling may be viewed as both intermicellar and intramicellar; in the kaolinite type of clays swelling is intermicellar only.

In addition to the kind of clay mineral involved, swelling is influenced by the nature of the adsorbed cations. Baver (1940: 80) indicates that the order of swelling of montmorillonite saturated with certain cations is as follows: $\text{Na} = \text{Li} > \text{K} > \text{Ca} = \text{Ba} > \text{H}$. Since the order evidently differs for other colloidal clays, it is not possible to generalize on this point. Likewise it is not clear why the various cations result in different amounts of swelling. Mattson (1932*b*) advanced the view that swelling is an osmotic phenomenon caused by the osmotic effect of the dissociated cations. He believed that a highly dissociated cation, such as Na, will cause greater swelling than a cation such as Ca, which is less readily dissociated. It is a fact that Na clays swell more than Ca clay; the volume occupied by 1 gram of Sharkey soil colloid after complete imbibition was as follows: untreated, 1.85 cc.; Ca-saturated, 1.9 cc; Na-saturated, 7.2 cc. Although osmotic effects are probably important in the swelling of colloidal clays, these effects alone seem insufficient to account for the variations in swelling which are known to occur. This subject requires much additional research before clarity can be achieved.

FLOCCULATION OF COLLOIDAL CLAYS

Flocculation is an important phenomenon from the standpoint of the physical properties of soils. The process may be readily demonstrated by adding a small amount of a neutral salt to a colloidal clay suspension. The individual clay particles coalesce, forming floccules which sink to the base of the liquid column. Although flocculation should be recognized as distinct from aggregate formation, it has a bearing on this process. It may be re-

garded as one of several conditions necessary for development of stable soil aggregates.

Clay colloids with low zeta potentials⁵ flocculate more readily than those with high potentials (Jenny and Reitemeier, 1935). In other words, colloidal suspensions tend to be unstable if the potential of the particles is low but incline toward stability if the potential is high. Particles in colloidal suspensions are in constant movement (Brownian movement) with the result that they are continually colliding. As long as the particles carry a strong charge, they exert a mutual repulsion, and after colliding move apart; under these circumstances the system undergoes no permanent change. When the potential falls below a certain critical level, however, the electrostatic repulsion is insufficient to prevent union of the particles, and flocculation occurs. The magnitude of the zeta potential is believed to increase with an increase in the number of charges on the clay crystal surface and in the average distance between the inner and outer ionic layers.

Considerations of hydration have been widely employed in explaining the influence of the size of adsorbed cations on the zeta potential. Hydration has been thought to increase with a decrease in the size of the ion, with the result that the smallest ions possess the thickest hull or shell of water. The thick hull of water carried by the highly hydrated ion is regarded as preventing a close approach to the surface of the clay crystal. Weakly hydrated ions, on the other hand, are able to approach closely. Recently, however, Grim (1942) and others have cast doubt on the commonly accepted views concerning hydration of cations. It now appears that certain cations, for example, Na^+ , previously thought to be highly hydrated are probably not hydrated at all and that others are less hydrated than has been commonly assumed. Clarification must await further research.

In general, the magnitude of the zeta potential of colloids saturated with monovalent cations is greater than that of colloids saturated with divalent cations. Evidently there are exceptions to this statement, but it applies reasonably well to cations of approximately the same size. Thus, the zeta potential of Li clay is higher than that of Mg clay, and similarly NH_4 clay has a higher potential than Ba clay.

⁵ The zeta potential is the charge on the colloidal particle and is usually negative.

The common bases in soils may be arranged in the following order of decreasing capacity to bring about flocculation of clay: $\text{Ca} > \text{Mg} > \text{K} > \text{Na}$. It may be noted that this is also the order of their usual strength of adsorption or replacing capacity. In other words, cations which are strongly adsorbed and difficult to replace favor flocculation, whereas cations that are weakly adsorbed and easy to replace favor dispersion. It is quite generally accepted that clays whose exchangeable cations consist principally of H^+ or Ca^{++} tend toward flocculation and that clays carrying Na^+ as the principal exchangeable cation incline toward dispersion. The explanation for the high flocculating capacity of hydrogen ions is not clear.

ORGANIC COLLOIDS

One of the first to recognize that humus possesses colloidal properties was van Bemmelen (1888). Some investigators regard only the finely divided, rather well-decomposed, organic matter as colloidal; but others, for example, Anderson *et al.* (1922) regard all the soil organic matter as colloidal, irrespective of its state of subdivision. This view seems to be justified, since plant debris usually possesses colloidal properties.

Organic soil colloids contain components soluble in ether, alcohol, and water and generally have considerable cellulose and hemicellulose. In a number of samples examined by Anderson and Byers (1933) the largest single component was the ligno-humic complex. The importance of lignin in soil humus has also been recognized by Marshall (1935: 105), Waksman (1938), and others.

Electrodialized organic colloids are strongly acid. In a number of samples investigated by Anderson and Byers (1933) the viscous material showed pH values ranging from 2.3 to 3.7. When free from electrolytes, humus particles are distinctly electro-negative and readily dispersible. Humus has the property of combining with bases to form humates and in this condition is less susceptible to dispersion (Mattson, 1931).

When equal weights are considered, the combining power of organic colloids is much greater than that of the mineral colloids, even when the mineral colloids are highly silicated. The data from Anderson and Byers (1933) in Table 32 illustrate this fact.

TABLE 32. SODIUM HYDROXIDE REQUIRED TO NEUTRALIZE ELECTRODIALYZED COLLOIDS AND THEIR AMMONIA-HOLDING CAPACITY

(After Anderson and Byers, 1933)

Type of Colloid	NaOH Required to Neutralize 1 gm. of Colloid, m.e.	NH ₃ Held by 1 gm. of Colloid, m.e.
Forest leaf mold	0.967	1.600
Cecil soil, H ₂ O ₂ treated	0.057	0.104
Fallon soil, H ₂ O ₂ treated	0.580	0.668

The Fallon colloid has a high silica-sesquioxide ratio and is capable of holding more base than are most inorganic colloids; the Cecil colloid has a low silica-sesquioxide ratio and holds only a fraction of the bases held by the Fallon colloid. The so-called forest leaf mold colloid was obtained from New Hampshire and was derived mainly from birch leaves. This colloid is shown to possess a materially higher base capacity than the Fallon colloid when equal weights are considered. It may be pointed out, however, that the apparent as well as the actual specific gravity of organic colloids is much lower than that of inorganic colloids. Therefore the exchange capacity of a given volume of soil composed largely of organic colloidal material is not necessarily higher than that of an equal volume of soil composed largely of inorganic colloidal material. Let us assume a hypothetical situation involving an organic soil with an exchange capacity of 250 m.e. per 100 grams and a soil composed entirely of inorganic colloids with an exchange capacity of 50 m.e. per 100 grams. Let us further assume that the volume weight of the organic soil is 0.25 and that of the inorganic soil is 1.25. These exchange capacities, after being corrected to equal volumes, are found to be identical. This hypothetical case must not be interpreted as indicating a general equivalence between the exchange capacities of inorganic and organic colloids when they are considered on the basis of equal volumes. Rather it is intended to illustrate the need of recognizing differences in density when comparing these exchange capacities.

Organic colloids are affected by the nature of the adsorbed cations in much the same way that inorganic colloids are affected. They are not readily flocculated when saturated with monovalent metallic cations. In nature, however, and especially in

forest soils, the dominant cations are hydrogen and calcium, with the result that the organic colloids tend to be flocculated.

The importance of organic colloids in forest soils is particularly great. Not only is the exchange capacity enhanced by incorporated organic matter but also there are other significant influences on the physical, chemical, and biological conditions. Certain of these effects have already been considered; the others will be discussed later.

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Chapter 8

GENERAL PHYSICAL PROPERTIES OF FOREST SOILS

Physical properties of the soil are highly important from the standpoint of both soil science and plant ecology. Even the biological and chemical characteristics are influenced by the physical properties. From time to time various investigators have placed particular emphasis on the importance of the physical characteristics of soils in relation to plant development.

Thurmann (1849, 1: 274) advanced the view that physical, not chemical, properties of soil were of primary importance in determining vegetation development. This concept, becoming widely accepted, exerted a powerful, and unfortunate, influence on the early development of plant ecology. Recognition of the importance of physical properties led many investigators to minimize or even ignore the influence of chemical properties. Among foresters who have regarded physical properties as of outstanding importance the following may be mentioned: Henry (1908: 266), Bühler (1918: 379), Burger (1924), Dengler (1935: 160), and Krauss and Härtel (1935).

It should be recognized that within limits excellence in physical properties of a soil may tend to compensate for poorness in chemical composition. The most favorable soil conditions however, require excellence in all factors. Soil productivity cannot be specified in terms of physical characteristics alone.

Certain physical properties of forest soils are susceptible to change and are, to some extent, subject to influence by the silviculturist; soil structure and porosity may be mentioned as examples. Other properties, such as texture, cannot be altered except by means that are not economically feasible. One of the objectives of the silviculturist should be the development and maintenance of the highest possible soil productivity in the areas under his management. Knowledge of the various physical properties of soils is a necessary prerequisite to the attainment of this objective.

THE SOIL PROFILE

A vertical section through the soil body, as seen in an excavation, usually exhibits a series of layers or *horizons* of varying thickness. This series of horizons in their natural sequence is called, collectively, the *soil profile*.

The horizons of a soil profile may differ greatly in physical, chemical, and biological characters but they are related genetically. In this respect they are unlike the strata of geological formations. Just as the horizons of a specific profile differ, so the horizons in unlike profiles are different. In view of the tremendous variability found in different profiles it is impossible here to anticipate all possibilities. However, it seems desirable to indicate in a broad way the principal features of what may be designated as a generalized or hypothetical profile (Figure 28). It is emphasized that the following remarks apply to a purely hypothetical soil profile and that the horizons indicated may or may not be found in specific soils in the field.

The principal horizons of the soil profile, in the order in which they appear from top to bottom of a vertical exposure, are usually designated by the letters *A*, *B*, and *C*. Each of these major horizons is commonly subdivided. The uppermost layer, consisting of fresh or only slightly altered organic material (litter), is designated as the A_{00} layer or the *L* layer. Immediately below is the A_0 layer, consisting of organic matter in a more advanced state of decomposition. Forest-soil workers commonly recognize two subdivisions of the A_0 layer. The upper part of the A_0 layer, consisting principally of partially decomposed organic matter which still retains sufficient structure to permit identification of its source, is designated the *F* layer; it is the upper portion of the humus layer.¹ The lower part of the A_0 layer, consisting principally of amorphous organic matter, is designated as the *H* layer. Thus in mor humus layer types the A_0 layer is made up of an *F* layer and an *H* layer. In mull humus layer types the situation is somewhat different, as is explained in Chapter 6. The A_0 and A_{00} layers may be absent in soils developed under grasses.

The *A* horizon, frequently subdivided into the A_1 , A_2 , and A_3

¹ The term *humus layer* is applied to the top portion of the soil, which owes its characteristic features to its content of humus. For descriptions of humus layer types refer to Chapter 6.

layers, includes the upper part of the mineral soil profile. From this region material has been removed in suspension or solution by percolating water. The A_3 horizon marks the lower boundary of the zone of eluviation. In normal soils of humid regions the

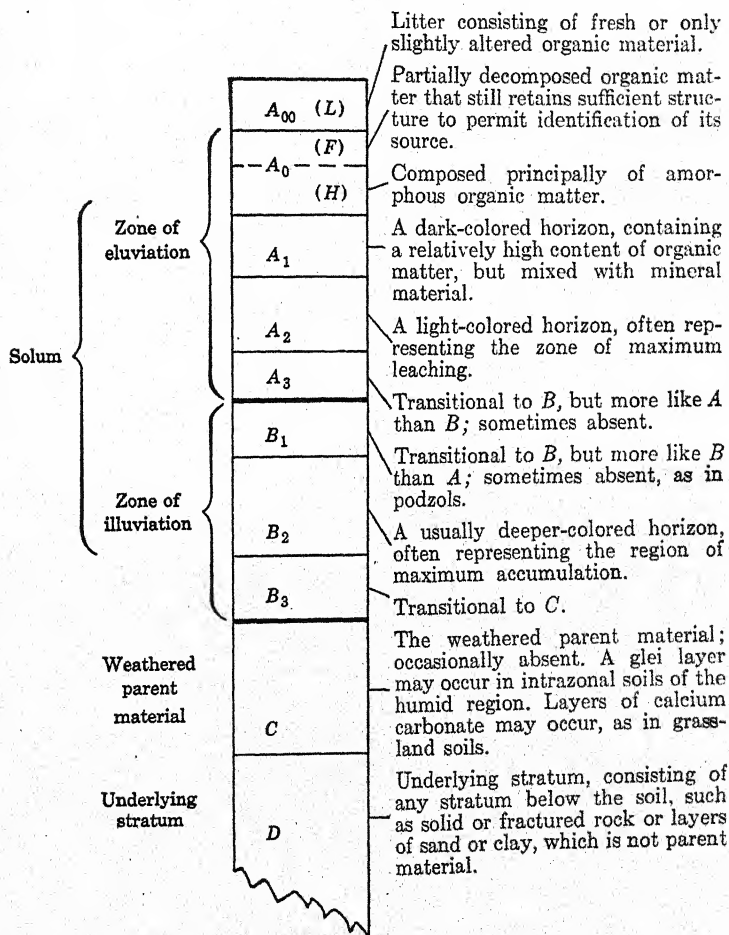


FIG. 28. A hypothetical soil profile having the principal horizons.

B horizon, frequently subdivided into B_1 , B_2 , and B_3 layers, is characterized by deeper colors and heavier texture. It is known as the zone of illuviation or accumulation. Underlying the B horizon is the weathered parent material, or C horizon. A more

detailed consideration of soil profiles and horizons will be found in Chapter 11.

THE SOIL PARTICLES

A soil may be regarded as a system of three phases: solid, liquid, and gaseous. The solid phase is represented by an intimate mixture of inorganic and organic matter. The inorganic constituents consist of rocks, minerals, and their decomposition products. In shape the particles are highly variable, and in size the material ranges from large boulders and cobbles to molecules and ions in solution. In this section only material of colloidal dimensions or larger will be considered. The fine particles may exist singly, as aggregates, or as coatings on the large particles. Many of the particles are porous. *Soil texture* refers to the relative proportion of the various size groups of individual particles; the individual size groups of soil particles are referred to as *soil separates*.

SIZE CLASSES OF SOIL PARTICLES

In view of the wide variation in size of individual particles of soil material it is necessary to recognize size classes or groups into which the particles may be separated for purposes of study and classification. Various size-class scales have been used in different parts of the world. One feature common to all scales is the arbitrary nature of the size classes employed.

Attention has been primarily devoted to the classification of particles less than 2.0 mm. in diameter. This material is usually known as fine earth. In 1913 an international commission adopted a classification referred to as the International or Atterberg System (Atterberg, 1912a). This classification is simple, and the successive particle-size classes are at equal logarithmic intervals. The size classes employed are presented in Table 33. The system most widely used in America is quite different, since it employs seven size classes for material less than 2.0 mm. instead of the four size classes of the Atterberg system. It will be noted in Table 33 that the present American system incorporates certain changes which were adopted in January, 1938. At the present time it is the practice of the United States Department of Agriculture to report mechanical analyses by the International System as well as the American System. The fraction less than

0.005 mm. is still reported, but only in order that comparisons may be made with earlier data. The popularity of the International System is increasing in the United States. Use of this system seems advisable because of its simplicity and because of the doubtful importance of the various sand separations recognized in the American system.

TABLE 33. SIZE CLASSIFICATIONS OF SOIL PARTICLES

International (Atterberg) Classification		American Classification	
Conventional Name of Separate	Size Class, Diameter Limits, mm.	Conventional Name of Separate	Size Class, Diameter Limits, mm.
Coarse sand	2.0 -0.2	Fine gravel	2.1 -1.0
Fine sand	0.2 -0.02	Coarse sand	1.0 -0.5
Silt	0.02-0.002	Medium sand	0.5 -0.25
Clay	< 0.002	Fine sand	0.25-0.10
		Very fine sand	0.10-0.05
		Silt*	0.05-0.002
		Clay*	< 0.002

* Before January 1, 1938, the fraction 0.05-0.005 mm. was called silt, and the fraction less than 0.005 mm. clay.

Very little attention has been paid in the past to the classification of soil material larger than about 2 mm. in diameter. In forest soils this larger-sized material may be very abundant and important. The classifications of Wentworth (cited by Twenhofel, 1939: 185), Atterberg (1908), and Ramann (1911) appear

TABLE 34. SIZE CLASSIFICATIONS OF SOIL MATERIAL
LARGER THAN 2.0 MM. IN DIAMETER

Wentworth System Size Class, Diameter Limits, mm.	Atterberg System* Size Class, Diameter Limits, mm.	Ramann System* Size Class, Diameter Limits, mm.
> 256.0 (boulders)	> 2000.0	> 40.0
256.0-64.0 (cobbles)	2000.0-600.0	40.0-2.0
64.0- 4.0 (pebbles)	600.0-200.0	
4.0- 2.0 (granules)	200.0- 60.0	
	60.0- 20.0	
	20.0- 6.0	
	6.0- 2.0	

* Names for material in the various size classes are not given because of the difficulty in choosing precise English equivalents for the German names employed by Atterberg and Ramann.

suitable for use by forest-soil workers. Size-class limits recognized in these systems are presented in Table 34. The Wentworth scale appears to be satisfactory for work with most soils; if a larger number of size classes are desired, the Atterberg system is worthy of consideration. In addition to classifying the material larger than 2 mm. into size classes, Ramann (1911) recognized two categories based on shape, namely, rounded material and angular material.

MECHANICAL ANALYSIS OF SOILS

A mechanical analysis of soil material involves the separation of the individual or primary particles into size classes and the determination of the percentage by weight in each class. The purpose of a mechanical analysis is to supply information on the general physical properties of a soil and to furnish the facts necessary for specification of the textural class. Mechanical analysis data define a fundamental and stable soil property; coarse sandy soils remain coarse sandy soils, and clay soils remain clay soils.

It is now generally accepted that one of the first essentials in making a mechanical analysis is complete dispersion of the soil particles, which in nature may occur in the form of aggregates. Cementing materials, such as calcium carbonate and organic colloids, must be removed.

Various methods of mechanical analysis are available. The hydrometer method, perfected by Bouyoucos (1936, 1937a, 1937b) has found widespread favor among investigators of forest soils and is being used in many laboratories. It has a number of advantages, among which simplicity, rapidity, and reasonable accuracy may be mentioned. The pipette method, although far less rapid, is used extensively in this country and abroad. Many variations of the pipette method have been introduced, but in America the most widely used technique is that of the United States Department of Agriculture, described by Olmstead, Alexander, and Middleton (1930) and Shaw and Miles (1940). Details of these methods may be found in the original papers or in Wright (1939). For information on other methods Odén (1925, 1926), Keen (1931), and Baver (1940) should be consulted.

Material larger than 2.0 mm. is usually not considered in a mechanical analysis of an agricultural soil. This fact does not

mean that information concerning the amount and size-class distribution of soil material larger than 2.0 mm. is unimportant. On the contrary, there is considerable evidence that rocks influence site quality, and it is well known that forest soils are frequently rocky. Material larger than 2 mm. but not over 20 mm. in diameter may be determined in samples collected for the usual mechanical analysis. However, it is generally impractical to transport to the laboratory samples of sufficient size to permit determination of the amount of material larger than 20 mm. in diameter. This larger material may be estimated or determined by weighing in the field. Size classes such as those given in Table 34 may be used in making the separations. The volume occupied by material larger than 2.0 mm. may be calculated from weight and specific gravity values.

TEXTURAL DESIGNATIONS

Even the most cursory examination makes apparent the fact that in certain soils relatively coarse particles, and in others very fine particles, predominate. This has led to recognition of broad soil-textural classes, such as sands, loams, and clays. Soils are classified according to the texture of the upper part of the mineral soil body. In the field the broader variations in texture may be detected by the "feel" of the moist soil between the fingers. Modifications of the feel of soil material due to the presence of organic matter and different amounts of moisture should be appreciated. Lyon and Buckman (1937: 41) have presented the following key for the field identification of mineral soils:²

- I. Soils possessing the properties of one size of particle largely
 - A. Particles of soil very large *Gravel*
 - B. Particles apparent to eye; soil feels gritty and is rather nonplastic *Sands*
 - C. Particles very small; soil very plastic when wet, hard when dry *Clays*
- II. Soils possessing the properties of a number of sizes of particles — a mixture
 - A. A fairly equal exhibit of sandy and clayey properties *Loam*
 - B. A mixture but with sand predominating *Sandy loam*

² From *The nature and properties of soils*, by T. L. Lyon and H. O. Buckman. By permission of The Macmillan Company, publishers.

- C. A mixture but with silty characters dominant;
soil having a floury or talc feel and quite plastic
when wet *Silt loam*
- D. A mixture but with clayey characters very ap-
parent; soil very plastic and approaching a clay in
character *Clay loam*

In view of the fact that the physical characteristics of a soil may be strongly influenced by the size of the component particles it is commonly desirable to have a more definite specification of their size-class distribution than can be obtained by inspection in the field. This information is furnished by a mechanical analysis and serves as a basis for designating the textural class to which the soil belongs.

The textural classification which has been used in the Department of Agriculture follows. Twenty principal and subordinate textural classes are recognized (Davis and Bennett, 1927), as follows:

- I. Soils containing less than 20 per cent clay
 - A. Soils containing less than 15 per cent silt and clay (sand)
 - Coarse sand* (35 per cent or more fine gravel and coarse sand, and less than 50 per cent fine or very fine sand)
 - Sand* (35 per cent or more fine gravel and coarse and medium sands, and less than 50 per cent fine or very fine sand)
 - Fine sand* (50 per cent or more fine and very fine sand)
 - Very fine sand* (50 per cent or more very fine sand)
 - B. Soils containing 15 to 20 per cent silt and clay (loamy sand)
 - Loamy coarse sand* (35 per cent or more fine gravel and coarse sand, and less than 35 per cent fine and very fine sand)
 - Loamy sand* (35 per cent or more fine gravel and coarse and medium sands, and less than 35 per cent fine and very fine sand)
 - Loamy fine sand* (35 per cent or more fine and very fine sands)
 - Loamy very fine sand* (35 per cent or more very fine sand)
 - C. Soils containing from 20 to 50 per cent silt and clay (sandy loam)
 - Coarse sandy loam* (45 per cent or more fine gravel and coarse sand)
 - Sandy loam* (25 per cent or more fine gravel and coarse and medium sands, and less than 35 per cent very fine sand)
 - Fine sandy loam* (50 per cent or more fine sand, or less than 25 per cent fine gravel and coarse and medium sand)
 - Very fine sandy loam* (35 per cent or more very fine sand)

D. Soils containing 50 per cent or more silt and clay (loam and silt loam)

Loam (less than 20 per cent clay, from 30 to 50 per cent silt, and from 30 to 50 per cent sand)

Silt loam (less than 20 per cent clay, 50 per cent or more silt, and less than 50 per cent sand)

II. Soils containing from 20 to 30 per cent clay (clay loam)

Sandy clay loam (less than 30 per cent silt, and from 50 to 80 per cent sand)

Clay loam (from 20 to 50 per cent silt, and from 20 to 50 per cent sand)

Silty clay loam (from 50 to 80 per cent silt, and less than 30 per cent sand)

III. Soils containing 30 per cent or more clay (clay)

Sandy clay (from 30 to 50 per cent clay, less than 20 per cent silt, and from 50 to 70 per cent sand)

Clay (30 per cent or more clay, less than 50 per cent silt, and less than 50 per cent sand)

Silty clay (from 30 to 50 per cent clay, from 50 to 70 per cent silt, and less than 20 per cent sand)

Soils containing an appreciable amount of gravel or stone may have the appropriate term added to the textural designation, for example, gravelly coarse sand or stony clay loam.

An equilateral triangle similar to the one shown in Figure 29 can be used to determine the principal textural classes from mechanical analyses. Use of this triangle may be illustrated by an example. Point *A* is plotted from the following mechanical analysis data:

Percentage sand (2.0-0.05 mm.)	63.0
Percentage silt (0.05-0.005 mm.)	24.0
Percentage clay (0.005 mm.)	13.0

The position of point *A* indicates that the soil is a sandy loam.

Various investigators have sought to develop indices of texture. Hardy (1928) derived an index of texture as follows:

Index of texture =

Moisture content at the point of stickiness — $\frac{\text{Percentage of sand}}{5}$

His values varied from 60 and above for very heavy clay soils to less than 10 for sands. Barrington (1929) developed a texture index from mechanical analysis data. The assumption was made that all particles were spheres, and relative surface areas for each

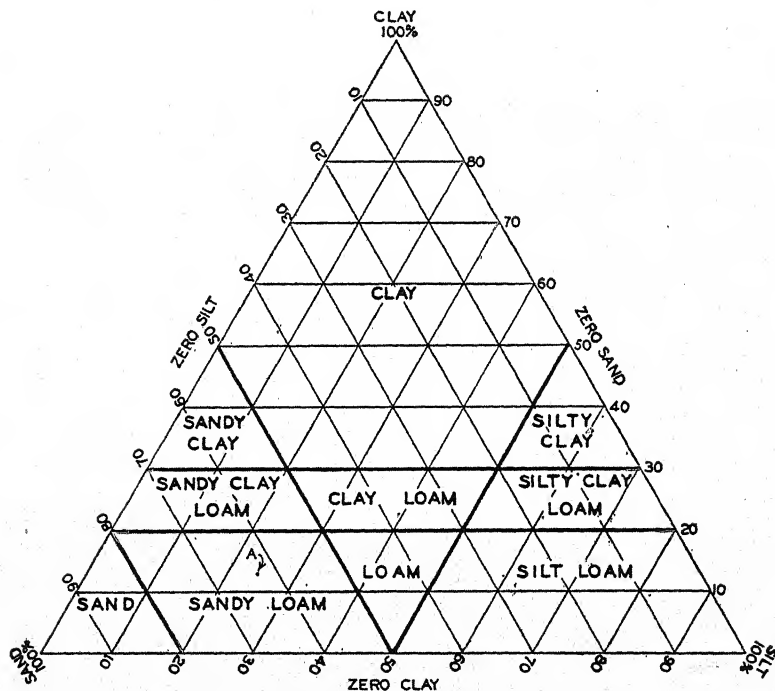


FIG. 29. A chart for determining the principal textural classes of soils.

fraction were computed. Soils containing large amounts of clay had high texture indices and sands had low indices.

Accurate specification of texture in the field is very difficult, if not impossible. Field textural designations are more of an art than a science; they seem to be based on a combination of features, of which particle size is only one. When doubt exists as to the proper textural designation, representative samples should be analyzed.

SURFACE AREA OF SOIL PARTICLES

It is well known that, as dispersion of material increases, the amount of surface also increases. Many of the important

physico-chemical phenomena of soils are influenced by the amount of surface per unit mass. The tremendous increase in surface area with increasing subdivision may be illustrated by assuming that a square mass of 1 cc. is successively subdivided as follows:

Length on Side, cm.	Number of Particles	Total Surface Area, sq. cm.
1.0	1	6
0.1	10^3	60
0.01	10^6	600
0.001	10^9	6,000
0.0001	10^{12}	60,000
0.00001	10^{15}	600,000

In this illustration the total surface area increases from 6 sq. cm. to 600,000 sq. cm. with subdivision of the initial cube of 1 cm. into particles 100 μ in size.

Baver (1940: 12) has presented calculations showing that in 0.5236 cc. of particles having a diameter of 0.05 mm. (50 μ) there is a surface area of 628.32 sq. cm. In the same volume of particles having a diameter of 0.0002 mm. (200 μ) there is a surface area of 157,080 sq. cm. The tremendous differences in the surface areas of coarse-textured and fine-textured materials help to explain many of the physico-chemical differences of soils.

PHYSICAL PROPERTIES OF SOIL SEPARATES

Gravel and coarse sand fractions tend to function as individual particles. The material is characterized by low cohesion, low plasticity, and low water-holding capacity. Soils in which these fractions predominate are apt to be dry because of low field capacity and excessive drainage. Aeration is usually good; the soils are "light" and, as a result of low field capacity, warm up rather early in the spring. Reasonable amounts of gravel and coarse sand are advantageous, particularly in soils containing much silt and clay.

Soils in which the fine sand and very fine sand fractions predominate are more cohesive and possess higher field capacity than those in which gravel and coarse sands predominate. Reasonable amounts of the fine sand fractions are desirable. Silt, because of the larger surface area it possesses, exhibits inter-

face phenomena to a greater extent than do the fine sands but to a lesser extent than clay and soil colloids.

Clay and soil colloids exhibit such properties as cohesion and adsorption to a marked degree. Soils composed largely of these fractions are "heavy"; they may be slowly drained, aeration may be inhibited, they may be difficult to penetrate, and they may be cold. During periods of dry weather such soils may shrink and crack.

PHYSICAL EFFECTS OF ROCKS

Since rocks are relatively good conductors of heat, their presence in a soil body is conducive to increased temperatures during the growing season (Grebe, 1886: 152; Wollny, 1897-98; Ramann, 1911: 294). From the standpoint of temperature relations rocks are advantageous in fine-textured soils, which may tend toward coldness because of high moisture content. Rocks also tend to make fine-textured soils more porous, so that penetration of water and air into the soil is facilitated (Senft, 1867: 269; Wollny, 1897-98; Ramann, 1911: 294; Bühler, 1918: 405; Bates, 1928). The presence of rocks on and in the soil surface also tends to decrease losses of water by evaporation. This fact has been reported by many investigators among whom may be mentioned Grebe (1886: 152), Wollny (1897-98), and Ramann (1911: 292). Long (1915) expressed the view that rocks inhibit losses of soil moisture by evaporation in the ponderosa pine region in the Southwest. He stated that in many places rocky soils have a good stand of young trees of all ages, whereas 40 or 50 yd. away, on soil devoid of rocks, no reproduction whatever can be found. This fact is illustrated in Figure 30.

It appears that, as a general rule, reasonable amounts of rock in heavy-textured soils are to be regarded as favorable for tree growth (Schumacher, 1864: 204-205; Grebe, 1886: 151; and Wiedemann, 1925: 61, 88). Pearson and Marsh (1935) have reported that gravel and stones in heavy clay soils in the Southwest favor reproduction of ponderosa pine. However, as the content of rock in a soil rises above about 20 per cent by volume, the unfavorable effects begin to outweigh the favorable ones. Root space becomes restricted, temperature extremes may be great, and the field capacity of the soil body is decreased. In

sandy soil rocks appear to have an unfavorable effect on plant growth.

ECOLOGICAL SIGNIFICANCE OF SOIL TEXTURE

Texture, or any other single soil characteristic, seldom determines site quality. However, it may be observed that, as a rule, loam soils are more favorable for forest growth than either coarse

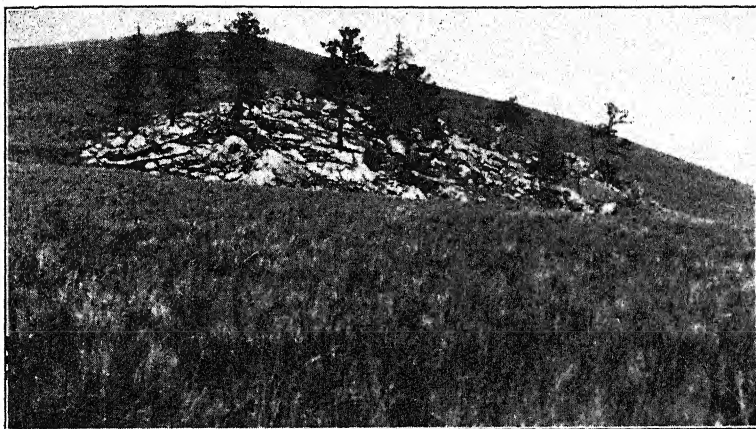


FIG. 30. Ponderosa pine on a rocky situation in a park within the ponderosa pine type, southwestern United States. The adjacent area supporting grass has a clay loam or clay soil. In the Southwest fine-textured soils are generally unfavorable for pine reproduction unless they are rocky. Photograph by J. O. Veatch.

sands or fine clays. Not infrequently the influence of texture on forest growth is masked by the influence of other, more significant, factors.

General observation indicates that deep, coarse sandy soils usually support relatively poor forest stands. Investigators have frequently reported that the content of material smaller than 0.2 mm. in sandy soils has an important bearing on site quality (Albert, 1925; Hartmann, 1926; Bornebusch, 1931). Other things being equal, the site quality of sandy soils increases as the proportion of material smaller than 0.2 mm. increases. The presence of layers containing fine-textured material in the lower part of a soil body may compensate for coarse-textured

material in the upper layers. Pearson and Marsh (1935) reported that regeneration and growth of ponderosa pine on deep volcanic cinder soils in the Southwest are commonly poor. However, cinder soils underlain with clay or loam produce good growth once tree roots have reached the heavier-textured layers.

Very heavy clay soils may also inhibit regeneration and subsequent growth of trees. Pearson and Marsh (1935) reported that the unfavorable features of clay soils in the Southwest are accentuated by trampling of livestock and by logging during wet weather.

In arid and semiarid regions fine-textured soils are conducive to the growth of grasses, whereas sandy soils favor trees (Henry, 1908; Stoeckeler and Bates, 1939). The more favorable conditions for tree growth on sandy soils in dry regions appear to be associated with larger amounts of available moisture.

SPECIFIC GRAVITY OF SOILS

The specific gravity of soil material is determined by the ratio between the weight of a given volume of solid soil particles and the weight of an equal volume of water. Specific gravity is unaffected by soil structure; in this respect it differs from volume weight.

In view of the wide variation of specific gravity observed in different soil minerals it might be expected that the specific gravity of mineral soil material would also vary between wide limits. Such variation, however, does not exist; it is generally accepted that the specific gravity of most mineral soil material free of organic matter is about 2.6–2.7. This is due to the fact that the minerals which make up the bulk of most soils have specific gravity values close to 2.65. The presence of large amounts of minerals containing iron results in specific gravity values above 2.6–2.7, and organic matter favors considerably lower values. Apparently the specific gravity of soil organic matter varies from about 1.2 to 1.7.

Specific gravity values are usually determined by means of a pycnometer. A first essential for accurate results is complete removal of all air from the sample; this procedure is rather difficult. Substantial errors occur, however, when measurements are made in cold water without pretreatment to remove air from the soil

material. Removal of entrapped air is conveniently accomplished by submerging the sample in water and subjecting it to a vacuum (Smith, 1943; Lutz, 1944a). Smith suggests the use of vacuum and atmospheric pressure alternately. Instead, most of the air can be expelled by boiling the soil in water for about 30 minutes (Burger, 1923). Good results have likewise been obtained by boiling the soil sample in xylol, a reflux condenser being used. If all air is not removed from the soil particles, the values obtained will be too low. Specific gravity calculations may be made as follows:

$$\begin{aligned}\text{Specific gravity} &= \frac{\text{Weight of dry substance in air}}{\text{Volume of substance}} \\ &= \frac{\text{Weight of dry substance in air}}{\text{Weight of dry substance in air} - \text{Weight of substance in water}}\end{aligned}$$

VOLUME WEIGHT OF SOILS

The volume weight of a soil may be defined as the ratio between the dry weight of a given volume of undisturbed soil and the weight of an equal volume of water. Thus, if 1000 cc. of undisturbed soil in dry condition weighs 1416 grams, its volume weight will be 1.416; in other words, it will be 1.416 times heavier than water. In view of the fact that volume weight, unlike specific gravity, is greatly influenced by soil structure, it is essential that the determination be based on the volume of soil in its natural field condition. Schumacher (1864: 123) was one of the first investigators to recognize this important fact.

For determinations of volume weight it is necessary to measure the volume of the soil with the included pore space, whereas in determinations of specific gravity the volume of the soil particles themselves, all pore space being excluded, is measured. Consequently, for a given soil, specific gravity values are always higher than volume weight values.

The volume weight of the A_0 horizon of forest soils is usually about 0.2 and that of the A_1 horizon is commonly less than 1.0, but considerably higher values, 1.5 or more, are obtained for samples from deeper horizons.

FACTORS AFFECTING THE VOLUME WEIGHT OF SOILS

One of the most important factors influencing the volume weight of soils is structure. Very compact soils, with low pore

volume, have high volume weights. Conversely, porous soils have low volume weights. The specific gravity of the constituent materials is also important. It is well known that, other factors being equal, soils with a high content of organic matter have a lower volume weight than soils with a low content of organic matter. The presence of rocks or sand in a soil favors high values, whereas the presence of fine fractions favors relatively low values. It will be apparent that, as a rule, volume weight increases with increasing depth below the surface. The increase of volume weight with greater depth is larger in forest soils than in grassland soils. The volume weight of a soil may also be influenced by the character of the vegetation which it supports. Investigators have reported lower volume weight values for surface soil from mixed forest stands than for soil from certain pure stands (Engler, 1919; Kmonitzek, 1930). Similarly, there is evidence that soil supporting grass sometimes has higher volume weight than comparable soil supporting forest. It appears unwise, however, on the basis of available information, to generalize as to the effect which various types of vegetation have on the volume weight of soils.

Excessive trampling of soil by grazing animals results in increased volume weight values, particularly when the soils are fine-textured and when the areas are used during wet weather. Intensive recreational use of forest park areas by the public likewise leads to compacting of the soil, as Burger (1940) and Lutz (1945) have shown.

ECOLOGICAL SIGNIFICANCE OF VOLUME WEIGHT OF SOILS

In soils which are inherently similar, low volume weight values signify a relatively porous condition, and high values indicate greater compactness. Poor aeration, slow infiltration of water, and relatively high resistance to penetration may be expected in fine-textured soils which have high volume weights. As a result tree root development will be inhibited.

MEASUREMENT OF VOLUME WEIGHT OF SOILS

A first essential in measuring volume weight is determination of the volume occupied by the soil samples *in situ*. It is impossible for the laboratory worker to restore in loose soil material

the physical conditions which existed in the field. The measurement of volume weight requires the collection of a sample of soil having a known volume and the determination of its dry weight. In soils relatively free from rocks metal cylinders have been employed to good advantage (Schumacher, 1864: 123; Burger, 1923; Auten, 1933; Coile, 1936; Lutz, 1944a). The size of cylinders used varies from as large as 2000 cc. to as small as 100 cc. Measurements made with small cylinders are usually less reliable than those made with large ones. In rocky soils where cylinders cannot be used, volume weight may be determined by the pit method recently developed by Lutz (1944b). This method has the advantage of permitting the use of large soil samples (1000-2000 cc.). The so-called paraffin immersion method (Trnka, 1914; Davis and Adams, 1928), employing relatively small blocks or lumps of soil, is also feasible when rocks prevent the use of cylinders. Among other techniques are the rubber tube method of Israelsen (1918) and the viscous fluid method of Beckett (1928). Neither of these two methods has come into general use.

Many soils, particularly those that are fine-textured, undergo an appreciable change in volume with a change in water content. Consequently, the moisture content at the time of sampling may influence volume weight values. It appears that whenever possible the moisture content of the soil should be at field capacity when sampling is carried out (Lutz, 1944a). Under this procedure the soil has an opportunity for full expansion under natural conditions before the samples are taken.

SOIL CONSISTENCE

The term *soil consistence* refers to the relative mutual attraction of the particles in the whole soil mass or their resistance to separation or deformation. This subject has been discussed in considerable detail by Atterberg (1912b), Keen (1931), and Baver (1940).

Terms commonly used to describe soil consistence are: *loose* or *open*, *slightly*, *moderately*, or *very compact*, *mellow*, *friable*, *crumbly*, *plastic*, *sticky*, *hard*, and *cemented*. Consistence is strongly influenced by soil moisture content; consequently it is necessary to indicate the moisture conditions at the time con-

sistence is defined. Cohesion and plasticity are both important in considerations of consistence.

SOIL COHESION

In moist soils cohesion is mainly due to surface tension effects produced by the water films distributed through the soil mass. When an annular ring or disc of water occurs at the point of contact of two particles, surface tension tends to draw the particles together. In dry soils cohesion is due to the attraction between the solid particles. Since cohesion varies inversely as the radius of the particles, fine-textured soils are more cohesive than coarse-textured soils. The total cohesion in moist soils, such as are usually encountered under field conditions, is the resultant of the temporary or induced cohesion due to water-film pressure and the true cohesion due to attraction between the soil particles.

The relatively large particles of sandy soils have low surface areas and in dry condition lack cohesion. Maximum cohesion values in sandy soils are attained at moisture contents which give maximum surface tension effects. At moisture contents above saturation, cohesion abruptly falls to zero or approaches it. Clays, in contrast to sands, attain their maximum cohesion when dry. The presence of films of water around clay particles results in decreased cohesion. The quality, as well as the quantity, of colloidal matter has a bearing on cohesion in soils. Organic colloids are less cohesive than inorganic colloids, and colloids having a low silica-sesquioxide ratio are less cohesive than those with a high ratio. The addition of organic matter to sandy soils tends to increase cohesion, whereas in clay soils it tends to decrease it. Haberlandt (1878) reported that the addition of 1 per cent of pulverized charcoal to a heavy-textured soil resulted in reducing cohesion about 36 per cent.

Methods for measuring cohesion are discussed by Keen (1931) and Baver (1940). In most investigations of forest soils, measurements of soil cohesion are not necessary.

SOIL PLASTICITY

The term *plasticity* refers to the ability of a body to change shape continuously under the influence of an applied force and

yet to retain the impressed shape on removal of the force. Plastic materials behave like rigid bodies, resisting deformation, when they are subject to forces less than a certain critical limit. If this critical limit is exceeded, however, flow occurs, and their behavior is then analogous to that of a highly viscous fluid.

Atterberg (1911) distinguished an upper and a lower plastic limit. The upper limit is the point at which the soil-water system or paste is just fluid. At this stage the system fails to retain the shape given it by an applied force. The lower limit is the point at which the system is no longer able to change its shape continuously under the influence of an applied force. At this point cylindrical samples crumble on being rolled or kneaded. The upper and lower limits of plasticity are expressed in terms of the moisture content of the soil, in percentage of dry weight; the difference between these two values is called the plasticity number. Highly plastic soils have relatively high plasticity numbers, and weakly plastic soils have low numbers.

Plasticity is strongly influenced by the quantity and quality of the inorganic colloidal material. Sandy soils, containing only small amounts of colloidal material, are nonplastic; fine-textured soils, containing relatively high amounts of colloidal material, are generally plastic. Clays with high silica-sesquioxide ratios are more plastic than those having low ratios. The cations present in the exchange complex also have an effect on plasticity; when clays are saturated with Na ions they are more plastic than when saturated with divalent ions, such as Mg or Ca. Particle structure is another factor influencing plasticity. Material composed of particles having a lamellar or sheet-like structure is extremely plastic. It is believed that, with adsorption of water films, the lamellar or plate-shaped clay crystals tend to become oriented, with the result that they can be made to move over each other readily. The water films between the oriented, plate-shaped clay crystals are regarded as holding them together and as having a lubricating effect.

Plasticity is not as important a consideration in forest soils as in agricultural soils, where cultivation is practiced. Nevertheless, even forest soils, if highly plastic, may be injured by disturbance during wet periods, for example, by trampling of grazing animals. When wet, the aggregates of a plastic soil are very easily broken down by trampling, with the result that the

soil, on drying, becomes hard and compact. The addition of organic matter reduces soil plasticity.

For a discussion of the measurement of plasticity in soil materials Atterberg (1911), Keen (1931), and Baver (1940) should be consulted.

SOIL STRUCTURE

Soil structure refers to the manner in which the soil particles are arranged. The ultimate particles of a soil, as determined by a mechanical analysis, usually do not exist in nature as independent units but are aggregated to form compound particles or particle groups. The development of stable aggregates is a complex process which involves the binding together of soil particles into structural units which are not dispersed in water. The quantity and character of these aggregates determine soil structure.

FACTORS INFLUENCING THE DEVELOPMENT OF AGGREGATES

Soil texture is one of the factors having a bearing on the development of aggregates. In very coarse sandy soils a structureless, or single grain, condition is often encountered; the possibility of aggregate development is small. On the other hand, clay soils show a wide variety of structural conditions, since the high content of colloidal material makes possible the development of aggregates of different types.

Organic matter is highly important from the standpoint of aggregate formation (Sokolovsky, 1934; Myers, 1937; Baver, 1940). The organic colloid particles bind the soil grains together, and it is believed that after dehydration relatively stable aggregates result. In most soils the higher is the content of organic matter, the greater is the degree of aggregation. Myers (1937) reported that organic colloids, either as hydrogen or calcium systems, were more effective in forming sand particles into water-stable aggregates than were clay colloids.

Aggregate formation is influenced by the cations present in the exchange complex. A widespread conviction has developed that calcium promotes granulation in soils and that soils rich in lime are usually in better physical condition than soils of the same texture which are poor in lime (Ramann, 1911: 302; Bühler, 1918:

379). This conviction appears to be based on two types of evidence: (1) direct observation of soils in the field, and (2) laboratory observations of the flocculation of colloid suspensions by calcium salts. It would appear, however, that the production of floccules and the production of water-stable aggregates is not the same. Recent work has shown that colloidal material saturated with calcium is not necessarily better aggregated than comparable material saturated with hydrogen; in fact, the reverse has sometimes been demonstrated (Bradfield, 1936; Bayer, 1940). Myers (1937) found that hydrogen systems, whether clay colloids or organic colloids, resulted in a greater degree of aggregation of sand particles than did the corresponding calcium systems. It seems probable that the favorable effects of calcium on soil structure which have been observed in the field are the indirect result of (1) increased activity on the part of the soil flora and fauna, and (2) the amount and nature of the soil organic matter.

Aggregates develop as a result of alternate wetting and drying of soils. Bayer (1940: 153) stated that "... alternate wetting and drying produce aggregation as a result of unequal strains and stresses that are set up by shrinkage and swelling processes, together with the disruptive action of air entrapped in the pores on wetting."³ It should be mentioned that aggregates formed as a result of alternate wetting and drying of soil are not very stable. Freezing and thawing are also known to favor the development of aggregates. Two principal reasons have been suggested for the formation of granules in soil which has been frozen: (1) pressure resulting from the development of ice crystals, and (2) dehydration of the colloids as a result of the withdrawal of water.

Ehrenberg (1922: 456) and others have reported that fires may result in the formation of aggregates in heavy clay soils. As a consequence of high temperatures reached during forest fires, small soil particles may be baked into aggregates which strongly resist deformation. It seems probable that part of the aggregating effect of fires is due to dehydration of the soil colloids.

Biotic agents play a highly important role in the development and maintenance of soil structure. Tree and other plant roots have frequently been credited with the formation of soil aggre-

³ Reprinted by permission from *Soil physics*, by L. D. Bayer, 1940, published by John Wiley and Sons, Inc.

gates. Pressures develop as roots grow, compressing soil particles into aggregates. It also has been suggested that the organic colloids resulting from root decay may contribute to aggregate formation. Further, as trees are swayed by winds, some of their roots may experience movement which results in loosening the soil. Soil fauna, particularly earthworms, millipedes, and other animals which devour mineral soil material, favor the development of crumb structure in the *A* horizon of many soils. Some workers have shown that slime and viscous microbial products encourage granulation.

FACTORS INFLUENCING THE MAINTENANCE OF SOIL STRUCTURE

As a general principle it may be stated that favorable soil structure is best maintained by healthy forest stands of species well adapted to their environment. Living plants and unincorporated organic matter resting on the mineral soil protect the aggregates from disruption by diminishing the violence with which rain drops strike. Consequently, exposure of the mineral soil for long periods should be avoided. The structure of clay soils is particularly susceptible to change as a result of prolonged exposure. Even periodic removal of litter from forest stands results in increased compactness of the soil (Ramann, 1898; Lunt, 1937). The harmful effects of this practice are probably due in part to the decrease of soil organic matter which accompanies litter removal.

Excessive trampling of the soil by grazing animals, particularly during wet weather, is injurious, since it results in increased compactness (Auten, 1933; Pearson and Marsh, 1935; Chandler, 1940). Damage is apt to be particularly noticeable on clay soils. The compacting of soil by visitors in forest areas subjected to intensive recreational use has sometimes created conditions unfavorable for tree growth (Meinecke, 1928; Lutz, 1945). Injury of this kind can be avoided, or at least greatly lessened, by careful planning of camp grounds and by application of special protective measures (Meinecke, 1932).

Soil cultivation of a superficial type is very occasionally employed in American forests to facilitate regeneration. In sandy and loamy soils cultivation usually results in more favorable

physical properties, but in clay soils the opposite effect may be produced. European workers have repeatedly observed unfavorable results from cultivation of heavy-textured forest soils (Lang, 1926: 440).

CLASSIFICATION OF SOIL STRUCTURE

Various classifications of soil structure and definitions of types of aggregates have been suggested from time to time (Zakharov, 1927; Kellogg, 1937; Nikiforoff, 1938). Kellogg has observed that all sorts of terms have been used for describing soil structure in the literature and that many of the terms are either poorly defined or not defined at all.

Some of the more important structural types with common modifications as presented by Kellogg (1937: 80-81) are:

Prismatic. Blocky structure with the vertical axis longer than the horizontal, as in the *B* horizon of many chestnut soils.

Nutlike. Blocky structure, angular as in the *B* horizon of the gray-brown podzolic soils, or rounded as in the *B* horizon of many chernozems.

Columnar. Prismatic with rounded tops, as in the *B* horizon of the solodized-solonetz.

Crumb. Generally soft, small, porous aggregates, irregular in shape, as in the *A*₁ horizon of many soils.

Granular. Hard or soft, but firm, small aggregates, angular or rounded, as in the *A* horizon of many chernozems.

Fragmental. Hard or soft, but firm, irregular aggregates, angular or subangular, as in many young soils developed from silty or clayey alluvium.

Phylliform. Thin layers, less distinct and thinner than platy. (When this condition is confined to inherited layers, as in the *C* horizon of soils developed from thin-bedded sediments, the term *laminated* should be used.)

To indicate a lack of definite structure, the following terms are normally used:

Single grain. Each grain by itself, as in dune sand (structureless).

Massive. Large uniform masses of cohesive soil, sometimes with irregular cleavage, as in the *C* horizons of many heavy clay soils (structureless).

The foregoing classification will be found adequate for most purposes. Kellogg suggested, "Each structural type should be modified appropriately so that the hardness, size, friability, and other characteristics of the aggregates are clearly defined. The approximate size should be given of prismatic, columnar, nutlike, and granular aggregates."

ECOLOGICAL SIGNIFICANCE OF SOIL STRUCTURE

From the standpoint of tree growth, structure is one of the most important physical properties of soil. Moisture relations and aeration conditions are both strongly influenced by soil structure; many other physical properties, such as pore volume, air capacity, volume weight, temperature, and permeability to water, are also influenced. Recognition of the ecological importance of soil structure, particularly in fine-textured soils, is general. In compact soils penetration of water is slow, aeration may be poor, root penetration is difficult, and the activity of soil animals is restricted. All these conditions are relatively unfavorable for tree growth. Occasionally in a forest the uppermost layer of mineral soil is compact, whereas the material in the lower layers is loose and aggregated. This situation indicates an unfavorable change or deterioration of the physical properties. The most desirable physical conditions are usually found in forest soils having a crumb structure.

The development of cracks in soils, as a result of shrinkage, may break plant roots and thus cause damage. Scholz (1935), in discussing the causes of decadence in old groves in North Dakota, suggested that the tree roots in the surface soil were injured by cracking of the soil. This type of injury has also been recognized by Ehrenberg (1922: 210). Dutta (1933) described another unfavorable effect associated with the formation of cracks in soil, particularly in areas where forest growth has been removed. The cracks which form tend to be wedge-shaped, that is, wider at the top than at the bottom. Heavy rains may wash material in, filling them before the soil body has had time to expand and cause their closure. According to Dutta, the "expansion joints" gradually disappear with years of exposure, and shrinkage without subsequent opportunity for expansion results in permanent reduction of porosity. Essentially the same situation has been

described by other investigators. Development of shrinkage cracks is most common in fine-textured soils which have become extremely dry.

AGGREGATE ANALYSIS AS A MEANS OF EVALUATING SOIL STRUCTURE

The objective of an aggregate analysis is to determine the extent to which the soil particles occur in the form of relatively stable aggregates. Aggregate analyses and mechanical analyses are similar in certain respects, but they differ in others. In both analyses the soil material is separated into a number of size classes, and the dry weight of each fraction is expressed as a percentage of the whole. The principal difference is that undispersed soil is employed in an aggregate analysis, whereas the soil is completely dispersed in a mechanical analysis. Numerous methods of aggregate analysis have been developed, but for present purposes they may be classified into three categories: (1) sieving, either dry or wet, (2) elutriation, and (3) sedimentation.

Russell and Tamhane (1940) discussed dry sieving of soils in the field. Their data showed that substantially different results were obtained by different workers sieving comparable soils. This method appears to have very limited value in studies of soil structure.

Wet sieving methods may be used to advantage in separating into size classes water-stable aggregates larger than about 0.25–0.50 mm. The soil sample, previously wetted by capillarity or by spraying with an atomizer, is placed in the top of a nest of sieves. The sieves are then immersed in water and shaken either manually or mechanically (Tiulin, 1933; Yoder, 1936). Mechanical shaking is superior to manual shaking, since mechanical treatment results in more nearly reproducible results. Meyer and Rennenkampff (1936) developed for aggregate analysis an apparatus employing an automatic siphon. Water is caused to rise slowly to the top of a nest of sieves and is then quickly removed by the automatic siphon.

Elutriation has been used in making separations of soil aggregates by various investigators (Demolon and Hénin, 1932; Baver and Rhoades, 1932). This technique is perhaps most useful in

separating aggregates of small size which cannot be classified by wet sieving.

Sedimentation techniques may also be used in separating aggregates of small size. The quantity of aggregates in suspension at any time may be determined by the pipette or hydrometer methods. Sedimentation tubes with segmented liners have also been successfully employed to determine the size distribution of water-stable soil aggregates. The tube designed by Moore and Kong (1941) has the advantage of permitting fairly rapid determinations.

Regardless of the technique employed in making an aggregate analysis, some of the material in each size class will consist of individual soil particles as well as aggregates. The situation in a given size class may be represented as follows:

Weight of material determined by aggregate analysis

= Weight of aggregates + Weight of soil particles existing as individuals

Consequently, the quantity of aggregates in a given size class is usually obtained by subtracting the weight of the soil particles, as determined by mechanical analysis of completely dispersed soil, from the weight of the material determined by aggregate analysis. Baver (1940: 172) has pointed out that an error is introduced when this technique is employed. For example, the weight of particles in the 1.0–0.5 mm. size class, as determined by mechanical analysis after complete dispersion of the soil, will not necessarily be the same as the weight of the particles which were associated with the aggregates in this same size class. "This error," Baver has indicated, "may be easily remedied if, instead of attempting to determine a size distribution of the completely dispersed sample, the aggregates on each screen (or in each tube of the elutriator) are dispersed and the material passed through the same sieve (or elutriating tube). The difference in weight before and after dispersion will more nearly give the amount of aggregates of that particular size."⁴

OTHER METHODS OF EVALUATING SOIL STRUCTURE

Various investigators have prepared sections of soil for microscopic examination of the structural features. A permanent

⁴ Reprinted by permission from *Soil physics*, by L. D. Baver, 1940, published by John Wiley and Sons, Inc.

record may be obtained by means of photomicrographs. For details concerning preparation of soil sections the works of Pigulevsky (1932), Harper and Volk (1937), and Kubiěna (1938) should be consulted.

Information on volume weight, pore volume, air capacity, and field capacity is helpful in evaluating soil structure. If it is assumed that texture and content of organic matter remain constant, the degree of aggregation in a soil may be expected to decrease as the volume weight increases. On the other hand, aggregation increases as pore volume and air capacity increase.

PORE VOLUME OF SOILS

The pore volume of a soil is the total volume not occupied by solid particles. In an absolutely dry soil the pores are completely occupied by air; in a water-logged soil the pores are occupied by water. The pores of forest soils normally contain both air and water, but the relative proportions are constantly changing. Soil pores may be visualized as consisting of a very complex system of dynamic cellular units or voids of irregular shape, which generally communicate with each other through narrow necks.

In considerations of pore volume it is convenient to employ the concept of capillary and noncapillary pores. Capillary pores are small and hold water by capillarity; noncapillary pores are large and do not hold water by capillarity. Capillary pores are primarily responsible for the capacity of soils to hold water against the force of gravity. Noncapillary pores are normally filled with air; they play an important role in soil aeration and internal drainage. Evaluation of the physical properties of soils, particularly as they relate to field capacity and infiltration of water, is greatly facilitated by information on the ratio between capillary and noncapillary pore volume. Two soils having identical total pore volumes may behave quite differently. For example, one soil may possess a low noncapillary pore volume and high capillary pore volume; this soil will have high field capacity, but infiltration of water will be slow. The other soil, with identical total pore volume, may possess a high noncapillary pore volume and low capillary pore volume; this soil will have low field capacity, but infiltration of water will be rapid.

Pore volume values of forest soils, as determined by Burger (1923), varied from about 28 to 69 per cent; values reported by Albert and Penschuck (1926) varied from 37 to 63 per cent. Baver (1940: 161) indicated that the porosity of average soils is in the neighborhood of 50 per cent. As a rule, pore volume decreases with increasing depth below the surface. The unincorporated organic matter of forest soils is very highly porous; porosity values of 80-90 per cent have been obtained.

FACTORS AFFECTING PORE VOLUME

Slichter (1899) investigated the nature of the pore space in systems consisting of spheres of uniform size. He established the fact that the pore volume between spheres packed as closely as possible is 25.95 per cent. In this arrangement each sphere touches surrounding spheres at twelve points. When the spheres are arranged less compactly, so that they touch one another at six points, the porosity is 47.64 per cent. King (1914: 117) recognized that, if the spheres are not of uniform size and if the smaller ones can pack the interstitial spaces between the larger ones, pore space may be reduced very markedly. He calculated that pore space may thus be reduced to 6.73 per cent for closest packing. King also calculated that the largest possible pore space is 72.58 per cent. These data are of interest, since they indicate the significance of arrangement and size of particles in considerations of pore volume. Pore volume increases with increasing aggregation of soil particles.

The amount and nature of the organic matter reaching the soil surface, together with the character and activity of the fauna and flora of the soil, are determined in part by forest stand composition. These factors in turn influence pore volume by affecting soil structure and content of organic matter. Indications are that the pore volume of soils supporting mixed stands may occasionally be somewhat higher than that of those supporting pure stands (Engler, 1919; Kmonitzek, 1930; Freise, 1934). The pore volume of soils which have been used for agriculture usually increases after reforestation.

Investigators in both America and Europe have shown that the pore volume of agricultural and pastured soils is lower than that of adjacent forest soils (Engler, 1919; Burger, 1923; Auten,

1933; Stewart, 1933). Auten found the porosity of field soils in the Ohio Valley to be 10-20 per cent less than in adjacent old-growth forests. The usually high porosity of forest soils is in large part due to their content of organic matter and to the presence of galleries formed by soil animals and canals left after the decay of roots.

Sometimes the pore volume of a forest soil increases as a result of exposure following clearcutting. Wittich (1930: 498) found this to be the situation in pure pine stands on sandy soil. The increased porosity appeared to result from decay of roots of herbaceous vegetation, which appeared after cutting. A similar tendency may be expected after removal of stands on podzol soils having a mor type of humus layer. On the other hand, there is evidence that the porosity of a soil, particularly if heavy-textured and if long exposed, may be reduced as a result of clearcutting (Wiedemann, 1925: 87; Gulisashvili and Stratonovitch, 1935). It appears that unfavorable results are confined to clearcut areas where development of a vegetation cover is delayed for a period of several years. Broad generalizations concerning the effect of clearcutting on soil porosity are unwarranted.

Other factors influencing the pore volume of forest soils are obvious. It is sufficient, therefore, to point out that any factor which is likely to decrease volume weight or increase aggregation will tend to increase pore volume.

ECOLOGICAL SIGNIFICANCE OF PORE VOLUME

The volume and nature of the pores in a soil are highly important from the standpoint of plant development. Total pore volume determines the amount of space available in the soil for air and water. The nature of the pores, whether capillary or noncapillary, determines field capacity, aeration, and internal drainage. Sandy soils are usually well aerated but have relatively low field capacity, because only a small proportion of their pore volume is in the form of capillary pores. Clay soils, however, usually have high field capacity but may be poorly aerated because of relatively low noncapillary pore volume. Probably the most favorable conditions are found in soils having a high pore volume about equally divided between capillary and noncapillary pore space.

Experience indicates that, other things being equal, soils of high pore volume are associated with good sites, whereas soils of low pore volume are associated with poor sites. Rohmeder (1937) found root rot in spruce due to *Trametes radiciperda* more prevalent in compact soils than in loose, well-aerated soils.

METHODS OF EVALUATING PORE VOLUME

In determining pore volume it is essential that the volume occupied by the soil *in situ* be known. Determinations made with pulverized soil samples, whose volume in undisturbed or field condition is unknown, are nearly valueless, since the natural architecture cannot be duplicated in the laboratory.

Total pore volume may be determined by either of the following formulas:

$$\begin{aligned}\text{Pore volume, per cent} &= \frac{\text{Volume of soil in place} - \text{Volume of solid soil particles}}{\text{Volume of soil in place}} \times 100 \\ &= 100 - \left(\frac{\text{Volume weight}}{\text{Specific gravity}} \times 100 \right)\end{aligned}$$

Methods of determining volume weight and specific gravity have been considered in earlier sections and need not be repeated here.

Measurements of total pore volume furnish no information concerning the nature of the soil pores. This information is supplied by measurements of the noncapillary and capillary pore volume. Noncapillary pore volume is equivalent to the air capacity; capillary pore volume is equivalent to the field capacity, expressed as a percentage of volume. Methods of determining air capacity and field capacity are discussed in later sections. For detailed information on methods of evaluating pore volume Kopecky (1914), Burger (1923), Stewart (1933), Gulisashvili and Stratonovitch (1935), and Lutz (1944a, 1944b) should be consulted.

AIR CAPACITY OF SOILS

Air capacity is usually defined as the volume of air in a soil having a moisture content equal to its field capacity. In other words, it is the volume of air in a moist soil after gravitational water has moved out. For practical purposes air capacity is

regarded as equivalent to noncapillary pore volume. Values are expressed as percentages of volume of undisturbed soil samples.

Data on air capacity indicate the volume of the noncapillary pore space but give little information concerning the nature or form of that space. Two soils having the same air capacity may differ appreciably if in one the noncapillary pores are intercommunicating and in the other they occur as individual voids separated by capillary pores. Interpretation of the nature of the noncapillary pores is facilitated by measurements of the rate of infiltration of water. High infiltration rates are associated with large, intercommunicating noncapillary pores. With increasing depth below the surface there is usually a decrease in air capacity.

FACTORS AFFECTING AIR CAPACITY

Air capacity, like other physical properties, is influenced by the vegetation which the soil supports. Investigators have reported higher air-capacity values for soils supporting mixed stands than for soils with certain pure stands (Němec and Kvapil, 1925; Kmonitzek, 1930). Canals resulting from the decay of roots in the *B* and *C* horizons contribute materially to the air capacity of these layers, as Engler (1919: 622) and others have pointed out. In this respect deep-rooted species play a more important role than shallow-rooted species (Figure 31). Cultural practices of the silviculturist, such as thinnings, also influence the physical properties of forest soils. Excessively dense stands are not favorable to high air capacity.

Heavy-textured soils that have been used for agriculture commonly have lower air capacity than adjacent forest soils, perhaps because of gradual filling of old root canals, breakdown of aggregates, and lessened activity by soil animals. Low air-capacity values for heavy-textured soils which had been cultivated or which occurred in pastures have been reported by Engler (1919) and others. The effects of agricultural use of coarse-textured soils appear to be quite different. In general, cultivation of sandy soils results in an improvement of their physical properties.

Air capacity is related to soil texture. Coarse sandy soils usually have high air capacities, whereas values for clay soils tend

to be relatively low, especially when the soil particles are not aggregated. The air capacity increases with greater size and

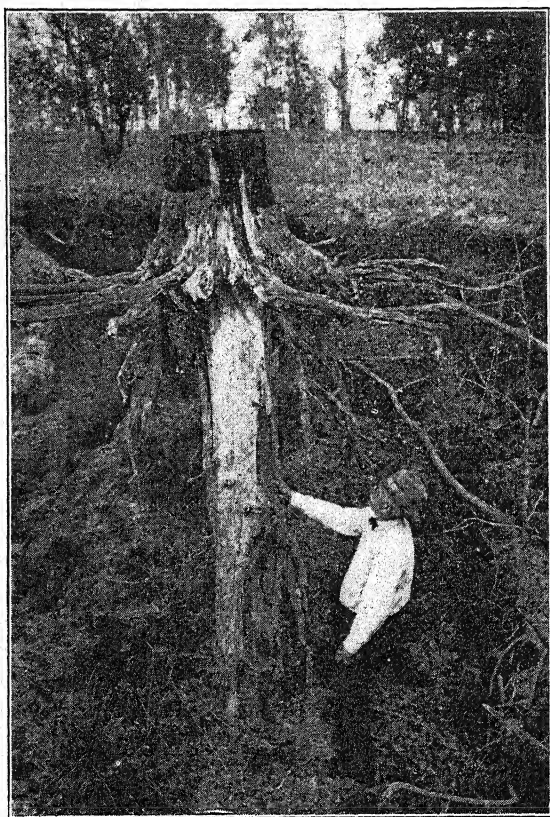


FIG. 31. The taproot of this longleaf pine stump in Mississippi has a diameter of 14 in. at a depth of 10 ft. below the ground surface. Photograph by United States Forest Service.

quantity of soil aggregates. The addition of organic matter to clay soils also increases air capacity.

ECOLOGICAL SIGNIFICANCE OF AIR CAPACITY

With certain notable exceptions, for example, in very coarse-textured soils, high air-capacity values generally indicate favorable physical conditions for tree growth. Soils having high air

capacity are usually well aerated, and infiltration of water is rapid. It should be recognized that for a given soil type the highest air-capacity values are not necessarily associated with the best sites, particularly in coarse-textured soils in which an excessive proportion of the total pore volume is in the form of noncapillary pores. Soils of this kind, because of a low content of capillary pores, have low field capacity and tend to be very dry. Here a reduction in air capacity and an increase in field capacity represent desirable changes. In clay soils a decrease

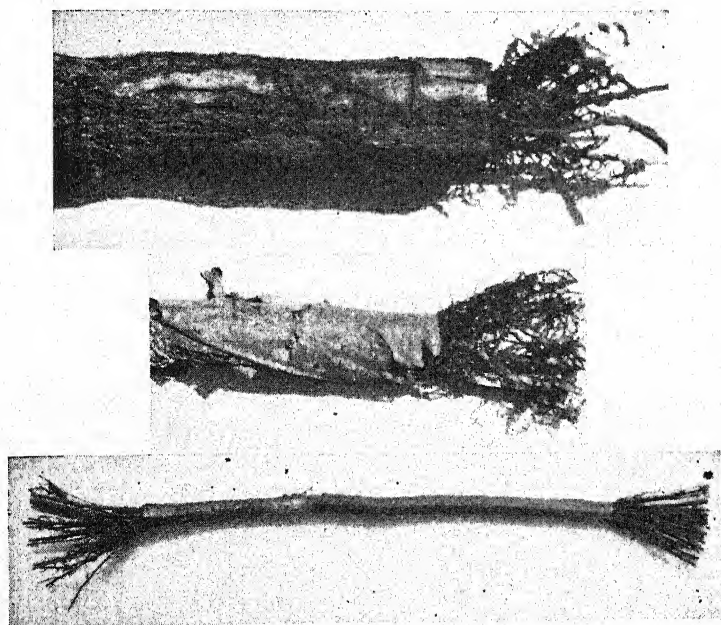


FIG. 32. In compact soils, living roots frequently are concentrated in old root canals. The examples shown above are about one-half natural size; only the bark of the dead roots remains (Yale Forest, near Keene, New Hampshire).

in air capacity usually signifies less favorable conditions for tree growth. Old root canals extending into the *B* and *C* horizons of heavy-textured soils favor deep penetration of new roots (Figure 32).

Burger (1924) observed that forest stands growing on heavy-textured soils which were formerly in agricultural use frequently showed poor development after they reached an age of 20-40

years. He attributed this poor growth to low air capacity and other unfavorable physical conditions of the old-field soil. Siegrist (1929) investigated two young spruce stands, one growing in soil having an air-capacity value of 10 per cent and the other in soil having an air-capacity value of 2 per cent. The stand growing in the soil of highest air capacity was thrifty; the needles were dark green and 12-20 mm. in length. The trees growing in the other soil were not thrifty, and the needles were yellow and only 4-6 mm. in length. Gulisashvili and Stratonovitch (1935) concluded that in the Leningrad region total and noncapillary porosity were the most important physical properties of forest soils.

MEASUREMENT OF AIR CAPACITY

In measurements of air capacity it is essential that the volume occupied by the soil *in situ* be known. The reasons for this requirement are obvious. The most widely used method for measuring air capacity has involved the use of samples collected in metal cylinders. The soil sample is first immersed in water with the object of replacing the air in the pores. After this step the system, supposedly containing only solid soil particles and water, is permitted to drain for 1 or more hours. The volume of gravitational water which moves out is regarded as equivalent to the volume of noncapillary pore space or the air capacity. There is little or no agreement among investigators as to the details of procedure. Sometimes a vacuum pump is used to facilitate replacement of the air in the soil pores by water; usually the sample is simply immersed in water for 24 or more hours. Furthermore, time allowed for drainage of gravitational water varies. Some investigators use fixed periods of 1 or 2 hours, but others determine the time for drainage of each soil type. For details of procedure the works of Burger (1923), Siegrist (1929), and Lüdi and Luzzatto (1935) should be consulted.

Two difficulties arise in connection with the method just outlined. First, the assumption that complete replacement of the air in the sample is obtained simply by immersing the soil in water for 24-48 hours is not justified. Actually, complete replacement by this treatment is impossible. Second, the amount

of water held in the soil sample after it is allowed to drain for 1 or 2 hours in the laboratory exceeds the moisture content at field capacity. The result is that air-capacity values determined by this method are too low.

Recently techniques have been developed which avoid these difficulties (Lutz 1944a, 1944b). These methods utilize samples collected either in metal cylinders or, if the soil is rocky, from pits. In either case the volume of the soil *in situ* must be ascertained. Before sampling, the soil is allowed to attain its field capacity under natural conditions, and its moisture content is determined. After removal to the laboratory the entire soil sample is permitted to stand in water for 24-48 hours and is then subjected to a vacuum for 2-3 hours, with occasional shaking, to remove the air. After this treatment the volume of the entire soil sample is determined in a large pycnometer. With these data the air capacity can be readily calculated.

AERATION OF SOILS

Soil aeration refers to the degree to which a soil is supplied with oxygen. In considerations of soil aeration it is convenient to view soil as a system consisting of solid, liquid, and gaseous phases. Most of the gaseous phase occurs as free air in the soil pores not occupied by water, but in addition gases are adsorbed on the surfaces of the soil particles and are dissolved in the liquid phase. Both the volume and composition of air in a soil are not static but undergo constant change. In general, the air content of a given soil varies inversely with the content of water.

The exchange of gases between the soil air and the atmospheric air is very largely the result of diffusion (Buckingham, 1904; Romell, 1922). Buckingham showed that the rate of diffusion increased with increase in volume of pore space not filled with water. Romell reported that the presence of water in soil pores decreased aeration through the plugged pores to about 1/10,000 of the normal amount.

COMPOSITION OF SOIL AIR

Soil air usually differs from atmospheric air in composition. It contains more carbon dioxide and less oxygen and is more

nearly saturated with water vapor than atmospheric air. Within the soil body oxygen is continually being utilized by plant roots and microorganisms; at the same time carbon dioxide is being liberated. Because of exchange between soil air and atmospheric air, carbon dioxide moves out of the soil, and oxygen moves in. This exchange must occur if aerobic conditions are to be maintained. The investigations of Romell (1922) led to the conclusion that the oxygen content of air in the upper layers of well-drained forest soils seldom falls much below that of atmospheric air.

The composition of the free air in a soil has frequently been regarded as a measure of aeration of the soil body as a whole. Various investigators have pointed out that this view is not justified (Romell, 1928: 78; Bayer, 1940: 265). Analyses of air extracted from a soil supply no information concerning the total amount of air in the soil or the degree of ramification of the air-filled pores in the soil body. Although the oxygen content of air extracted from two soils may be identical, in one anaerobic conditions may prevail in most of the soil mass, whereas in the other aerobic conditions may prevail.

FACTORS AFFECTING COMPOSITION OF SOIL AIR

Aeration is usually good in soils with a large volume of non-capillary pores forming an intercommunicating system well distributed throughout the mass. Carbon dioxide is able to move out of the soil air, and oxygen moves in. Old root canals and animal tunnels contribute very materially to the development of relatively large noncapillary pores. A deficit of oxygen and an excess of carbon dioxide frequently occur in very compact soils, especially during periods of high moisture content. The soil pores become filled with water, and gas exchange is then slow.

The presence of compact layers having low permeability inhibit aeration, as Ramann (1911) and Romell (1922) have pointed out. Romell investigated the composition of the air in a number of soils having hardpan; the highest carbon dioxide content found was 2.8 per cent, and the lowest oxygen content was 15.6 per cent, by volume. These samples were taken from a humus hardpan at a depth of about 50 cm. It should be pointed

out, however, that Romell also found oxygen contents of nearly 20 per cent in iron hardpans.

As a rule, the content of oxygen drops, and the content of carbon dioxide rises, with increasing depth below the soil surface. The change in composition of the soil air with increasing depth is less rapid in porous than in compact soils.

Poorly drained soils are very apt to be poorly aerated. The highest concentrations of carbon dioxide reported by Romell (1922) were about 5-6 per cent and the lowest oxygen values 1-2 per cent, by volume; these values were obtained in spruce forests, where the soil was very wet. However, oxygen is not necessarily deficient in all wet soils. Hesselman (1910), Clements (1921), and Romell (1922) have pointed out that the oxygen content of moving water is usually high. Tamm (1925) investigated the ground waters in morainic soils in northern Sweden and found, in general, a high content of oxygen. On the other hand, stagnant water is usually low in oxygen.

The content of carbon dioxide in the soil air is strongly influenced by the activity of microorganisms and plant roots. Russell and Appleyard (1915) attributed most of the fluctuations in the content of carbon dioxide in the soil air to fluctuations in the activities of microorganisms. Romell (1922) gave 5.2 grams (2.8 liters) per square meter per day as the production of carbon dioxide in certain Swedish beech forests.

Mor humus layer types have frequently been blamed for causing poor soil aeration. Assumptions rather than experimental data usually constitute the basis for claims of this kind. Mor types of humus layer may have some influence on gas exchange, but the investigations of Romell (1922, 1928) demonstrated that poor aeration is not a general result. In wet, poorly aerated soils mor types of humus layer may develop; here the humus layers are the effect, not the cause, of deficient aeration.

Ramann (1911: 386) and Russell and Appleyard (1915) believed that the composition of soil air is influenced by changes in temperature and barometric pressure, by penetrating water, and by the pressure and suction effects of winds. Romell (1922) also considered these factors but concluded that their influence was small. Diffusion is far more important.

The composition of soil air varies considerably during the course of a year. In general, investigators have found a minimum con-

tent of oxygen and a maximum content of carbon dioxide in mid-summer or late summer. During winter the situation is reversed; that is, maximum oxygen and minimum carbon dioxide values are recorded.

ECOLOGICAL SIGNIFICANCE OF SOIL AERATION

The ecological significance of an oxygen deficit and a carbon dioxide excess in soils has been summarized by Romell (1922: 343-344) as follows:

1. There are very great differences in the tolerance of different plants to an oxygen deficit and a carbon dioxide excess.
2. An excess of carbon dioxide or a deficit of oxygen of less than about 2 per cent is generally not harmful, even if maintained for a long period of time.
3. A complete or very pronounced restriction of gas exchange between the roots and the atmosphere, even if of relatively short duration, may result in disastrous effects in sensitive plants.
4. A moderate restriction of gas exchange, which in itself is not harmful, if long continued may result in the accumulation of toxic materials which cause damage.
5. Everything considered, it is apparent that under certain circumstances soil aeration may become an important ecological factor.

Research in plant physiology and agronomy has demonstrated that nutrient salt absorption by various plants is greatly favored by a good supply of oxygen to the absorbing roots (Broyer and Hoagland, 1943; Hoagland, 1944; Hoffer, 1945). Not only is the absorption of nutrients increased by good aeration, but absorption of water is also favored.

Deficient soil aeration has repeatedly been held responsible for poor development of forest trees. The evidence, if any, supplied to support this claim is often questionable. Too frequently poor development of a forest stand is attributed to aeration conditions simply because no other explanation is apparent.

MEASUREMENT OF SOIL AERATION

The measure of soil aeration usually employed is the composition of the soil air. It has already been pointed out that this

is an unsatisfactory index of aeration, but up to the present time no really satisfactory method has been available. Methods for obtaining samples of soil air and for their analysis have been described by Romell (1922, 1928) and Lundegårdh (1926).

Very recently Hoffer (1945) has developed a simple technique for diagnosing oxygen deficiencies in soils of corn fields. This technique is designed for field use and appears to hold great promise for the evaluation of oxygen relations in forest soils. The method is based on the fact that in most soils there are substantial amounts of iron and that these iron compounds are very responsive to aeration conditions. If aeration is good, most of the iron occurs in the ferric state, whereas under conditions of poor aeration most of the iron exists in the ferrous state. Hoffer's method enables the investigator to determine the relative extent to which ferric iron has been reduced to ferrous iron at different levels in the soil profile. For details the original publication should be consulted.

PENETRABILITY OF SOILS

The penetrability of a soil is the resistance offered to penetration by a solid body. Information on this property furnishes an indication of the compactness of soils; resistance to penetration increases as volume weight increases. The penetrometer measures a complex of soil conditions—resistance of the particles to separation, resistance of the soil to compression, and friction of the particles on the instrument.

The penetrability of a soil is influenced by particle size, structure, and moisture content. With increasing fineness of particles penetrability decreases, that is, resistance to penetration increases (Puchner, 1889). Moisture content and structure remaining the same, sandy soils are more easily penetrable than clay soils. Of course, the presence of rocks and large roots in a soil greatly affects measurements. Penetration is usually easier in soils having a crumb structure than in those of single grain structure. In fine-textured soils penetration becomes easier as the moisture content increases. Puchner (1889) reported that resistance in very dry or very wet sands was less than in sands of moderate moisture content. As a rule resistance to penetration increases with increasing depth below the soil surface.

Penetrability measurements have been found useful in evaluating the influence of different forest types on the physical properties of soils (Albert and Penschuck, 1926; Kmonitzek, 1930; Penschuck, 1931). They have also been used by Keen and Cashen (1932) in studies of the compacting effect of grazing animals and by Wahlenberg *et al.* (1939) in their investigation of the effects of fire and of cattle-grazing on longleaf pine lands.

MEASUREMENT OF SOIL PENETRABILITY

The penetrability of forest soils varies considerably from point to point. Consequently large numbers of measurements are usually necessary. In order to obtain an adequate body of data the penetrometer should be adapted for field use; it should be readily portable and should permit rapid measurements. Many different instruments meeting these requirements have been developed for specific investigations. The instrument most widely used is undoubtedly that developed by the Swiss engineer, von Meyenburg (1923). Values are obtained in terms of the number of kilograms of pressure necessary to force a metal body into the soil. Other types of instruments have been developed by Puchner (1889), Keen and Cashen (1932), Hénin (1936), and Shaw *et al.* (1943).

SOIL DEPTH

The thickness of the soil layers in which tree roots are able to develop freely influences forest composition and growth. Trees growing in shallow soils are usually less well supplied with water and mineral nutrients than those growing in deep soils. Losses resulting from the uprooting of trees are apt to be substantial in stands growing in thin soils where root development has been restricted.

Absolute, or physical, depth and physiological depth of soils may be distinguished. A soil may be deep in an absolute sense but, because of a relatively impervious layer, such as hardpan, or because of a high water-table, may be shallow in a physiological sense. Many swamp soils are physiologically shallow because of a high water-table which inhibits normal root development.

As a result of compensating conditions even shallow soils may support excellent forest stands. Seepage into a shallow soil may supply the vegetation with water and sometimes substantial amounts of mineral nutrients. Likewise, if the rock strata underlying a soil are fractured, tree roots may penetrate to depths of as much as 30 ft. When evaluating soil depth, the forester should always consider the condition of the underlying strata.

Classifications of soils from the standpoint of depth have been suggested by various foresters. Here is the classification which perhaps is most widely used:

Designation	Depth	
	Meters	Inches
Very shallow	< 0.15	< 6
Shallow	0.15-0.3	6-12
Moderately deep	0.3 -0.6	12-24
Deep	0.6 -1.2	24-48
Very deep	> 1.2	> 48

Strictly speaking, soil depth is the thickness of the combined *A* and *B* horizons, excluding the *C* horizon or parent material. However, because of the deep-rooted nature of most trees, the forester may find it convenient to employ an additional concept of soil depth which includes the parent material above the underlying strata.

SOIL COLOR

Of all soil characteristics color is perhaps the most obvious. For centuries farmers have used soil color as an index of soil fertility. It is an important aid in differentiating soil horizons and in classifying soils. Color has been used as the basis for naming certain great soil groups; for example, the *Braunerde* (brown earth) of the Germans, the red and yellow soils of the Americans, and the chernozem, or black soil, of the Russians. Color in itself is of minor significance, but it frequently serves as an indicator of other soil conditions which are extremely important.

FACTORS INFLUENCING SOIL COLOR

The color of most soils is due to pedogenic processes, but sometimes it is inherited from the parent material. The reddish

or red soils developed from red Triassic sandstones and shales in the North Atlantic States furnish an example of the inheritance of color.

The bulk of the mineral matter in soils, consisting of quartz and feldspars, is light-colored. Soil color is generally due to the presence of relatively small amounts of colored materials, such as iron and manganese compounds and organic matter.

Iron compounds in various stages of oxidation and hydration are extremely common coloring agents. Dorsey (1926) has indicated the color of certain iron compounds as follows:

Iron compound	Color
Ferrous oxide (FeO)	Green
Ferric oxide ($\text{Fe}_2\text{O}_3 \cdot n\text{H}_2\text{O}$)	
Limonite ($2\text{Fe}_2\text{O}_3 \cdot 3\text{H}_2\text{O}$)	Yellow, brown
Xanthosiderite ($\text{Fe}_2\text{O}_3 \cdot 2\text{H}_2\text{O}$)	Golden brown to brown
Göthite ($\text{Fe}_2\text{O}_3 \cdot \text{H}_2\text{O}$)	Yellow, brown, brownish black
Turgite ($2\text{Fe}_2\text{O}_3 \cdot \text{H}_2\text{O}$)	Red to reddish brown
Hematite (Fe_2O_3)	Red

It will be seen that in the ferric compounds, red colors are associated with low hydration and yellow colors with the higher stages of hydration. It may also be noted that there is a difference in color between the ferrous and ferric compounds. The colors produced by ferrous iron are less conspicuous than those due to ferric iron. Ramann (1911: 318) has pointed out that the color of a soil is not a reliable index of the amount of iron present. The color produced by 1-2 per cent of ferric oxide in a sandy soil may be about the same as that in loam and clay soils containing 5-10 per cent ferric oxide. Sands have relatively low total surface areas and require much less coloring matter to produce a given effect than is needed in clays, which have high total surface areas. The most vivid red colors are probably found in soils of the humid tropics. Ferrous compounds, such as pyrite, marcasite, vivianite, and glauconite, have been credited with producing the blue and green colors occasionally found in poorly aerated subsoils. Manganese compounds and magnetite are known to impart a dark color to soils; in this respect their effect is similar to that of organic matter.

Soil organic matter is dark-colored and is the most common cause of black or brown colors in soils. As plant remains undergo

decomposition, the color tends to become darker. Brown colors predominate in slightly decomposed plant materials, but the amorphous organic matter of soils is black or nearly so. The color imparted by organic matter is influenced by the texture of the soil. Sandy soils containing 0.2–0.5 per cent of humus may appear gray to black, whereas 5–15 per cent of humus is required in clay soils to produce the same degree of darkness.

If substantial quantities of organic matter are present, the color of the inorganic phase of the soil may be masked. The mineral color may be observed after destruction of the organic matter by boiling in hydrogen peroxide.

It is well-recognized that moisture content markedly affects soil color. Depth of color increases with increasing moisture, at least up to a certain point. Dark-colored soils seem to show greater changes in color with changes in moisture content than do light-colored soils. Robinson (1936: 192) suggested that color changes with variation in moisture content are greatest in soils containing a high percentage of colloids. In America descriptions of the color of soils usually refer to air-dry samples.

SIGNIFICANCE OF SOIL COLOR

Color is commonly regarded as a useful index of the degree and depth of weathering of the soil material. It is also indicative of drainage conditions, which, in turn, are related to soil aeration. In the presence of oxygen, ferrous oxide is unstable; it is oxidized to ferric compounds. Ferric compounds are usually darker colored than ferrous oxide. Red colors of soils usually result from the presence of unhydrated iron oxides.

Red colors generally indicate good soil drainage and aeration (Hilgard, 1906; Rice and Alexander, 1938). It does not follow, however, that all well-drained and well-aerated soils are red. Contrary to the opinion held by some persons, red soils are not characteristic of deserts but instead usually occur in regions having warm to hot, moist climates.

Yellow colors, due to the presence of hydrated iron oxides, are frequently found in imperfectly drained soils. The yellow color of some soils at present well-drained is believed to have been inherited from a previous condition of restricted drainage (Rice and Alexander, 1938).

Gray colors may result from poor aeration or low contents of iron and organic matter. The glei horizons of poorly drained soils are gray because of the presence of ferrous compounds, and the A_2 horizons of podzols are gray because of the leaching of iron and organic matter. The fertility of light gray or white soils in humid regions is usually low.

A specific soil horizon may be uniform in color, or it may be spotted or mottled. Mottling results from local accumulations of iron compounds, organic colloids, or other colored materials. It may also result from localized oxidation of iron compounds, such as frequently occurs in imperfectly drained subsoils. The importance of mottling as an indicator of poor drainage conditions is generally recognized.

DESCRIPTION AND MEASUREMENT OF SOIL COLOR

In describing a soil horizon, it should be noted whether the color is uniform or mottled. Aggregates, if any, should be crushed to determine whether the internal color is the same as the external. The gray appearance of some aggregates results from the presence of a thin coating of very finely divided silica. Granules of yellow or brown soil material may appear black because of a surface film of organic colloids.

Designation of the color of soils in the field at variable moisture contents is unsatisfactory. Whenever possible, air-dried samples should be used. Soil colors commonly recognized are black, gray, brown, red, and yellow (Kellogg, 1937). These principal colors may be modified to denote shade or depth. Thus, the term *reddish brown* or *yellowish red* designates a particular shade. In designating depth of color *dark gray* or *light brown* may be used. Modifying terms for both shade and depth are possible, for example, *dark grayish yellow* or *light reddish brown*.

A large number of schemes for the specification of soil color have been proposed. Color charts and scales, as well as powdered or granular materials, have been employed but have not proved entirely satisfactory. In 1928 Hutton described a method for obtaining rather precise measurements of soil color. This method, with or without the modification of Shaw (1932), has perhaps been more widely used than any other. Four standard color disks, with segments of each exposed, are rapidly rotated, and

the composite color compared with that of the soil. Adjustments are made until the color of the rapidly rotating disks exactly matches that of the soil. Color is then specified in terms of the percentage of white, black, yellow, and red. Lists of soil colors and their average factorial composition, as determined by the color-disk method, have been presented by Hutton (1928) and Kellogg (1937: 85). More recently Rice *et al.* (1941) have presented a set of preliminary color standards and color names for soils based on the Munsell color system. A color chart is included which permits direct comparison with any soil. The chart needs additional colors, but otherwise is quite satisfactory.

SOIL TEMPERATURE

Temperature is a highly important factor, since it influences, to greater or lesser degree, the physical, chemical, and biological processes in the soil. Soil temperature is determined by the balance between heat gained and heat lost. Heat gain relates to the absorption of direct solar radiation, absorption from warm air, and infiltration of warm rain water. Heat loss relates to radiation, conduction, and convection.

The principal source of heat is solar radiation. Other sources, such as conduction from the interior of the earth and chemical and biological processes in the soil body, are unimportant. Radiant energy reaching a given spot on the earth's surface is not constant; variations of diurnal and seasonal nature are well known. Furthermore, the intensity of radiation is greatly influenced by variations in the transparency of the atmosphere. A number of additional factors, some external and some internal, have a bearing on soil temperature.

EXTERNAL FACTORS INFLUENCING SOIL TEMPERATURE

Of the external factors which merit consideration, latitude and altitude, exposure and slope, and the living and nonliving cover of the soil may be mentioned.

LATITUDE AND ALTITUDE

Soil temperature decreases from regions of low latitude to regions of high latitude. Likewise, temperature usually de-

creases from stations of low altitude to stations of high altitude. The decrease in soil temperature with increasing altitude is seldom as rapid, however, as the decrease in air temperature. Fitton and Brooks (1931) pointed out that the annual range in soil temperature decreases with increasing altitude. Shreve (1924) found greater differences in maximum soil temperatures than in minimum soil temperatures at different elevations. On Santa Catalina Mountain, near Tucson, Arizona, the average maximum soil temperature at a depth of 3 in. on a north slope of 7000-ft. elevation was 27° – 29° F. higher than the average maximum at an elevation of 8000 ft. However, minimum temperatures reported by Shreve at the 7000-ft. elevation were actually lower than at the 8000- and 9000-ft. elevations. This situation was ascribed to strong nocturnal radiation at the 7000-ft. elevation, which was made possible by the relatively bare soil surface; at the higher elevations heat was conserved by organic debris covering the mineral soil. Nocturnal temperature inversions are not uncommon in mountainous regions (Hayes, 1941).

Peattie (1936: 18) and others have pointed out that as a rule the greater is the mass of a mountain, the higher is the mean temperature at a given elevation. Thus, forest limits are higher in the massive Alps than in the Fore-Alps of France.

EXPOSURE AND SLOPE

Exposure and degree of slope influence temperature, because they determine to a considerable extent the amount of solar radiation reaching the soil surface. The radiation received per unit area of surface is proportional to the cosine of the angle formed between a perpendicular to the surface and the direction from which the radiation comes. As this angle increases, the amount of radiation received per unit area decreases.

Soil temperatures on slopes having south and west exposures in the northern hemisphere are higher than on slopes having north or east exposures. Exposure has a greater influence on maximum temperatures than on minimum temperatures. The greatest differences are usually found when maximum temperatures on steep slopes having north and south or northeast and southwest exposures are compared. Temperatures are not only higher on south than on north slopes but also the range in temperature is

greater (Fitton and Brooks, 1931). During warm weather the air temperature on north slopes may be higher than the surface soil temperature. The reverse situation is found on south slopes.

Maximum surface soil temperatures attained on slopes having a south exposure are known to be 20°–25° F., or more, higher than on north slopes; mean differences in maximum temperatures on south and north slopes in southern New Hampshire during the summer have been found to be 13°–16° F. (Lutz, 1940). Alter (1913) stated that ground sloping 1° to the north lies in the same solar climate as level ground 70 miles further north. Likewise, land in southern Idaho which slopes 5° to the south is in the same solar climate as level land in the latitude of southern Utah, 350 miles nearer the equator. As inclination increases, north slopes become cooler and south slopes warmer.

LIVING COVER OF THE SOIL

The general tendency of living vegetation is to equalize soil temperatures; diurnal and seasonal variations are reduced. Because of interception of radiation the amount of insolation reaching the soil during the summer is decreased. On the other hand, vegetation also reduces the rate of loss of heat from the soil during the winter.

Vegetation lowers the maximum temperatures during the summer and sometimes raises minimum temperatures during the winter. The influence on maximum temperatures is much greater than on minimum temperatures. Differences in maximum surface soil temperatures in a forest and in the open are especially marked on clear days during the summer. The higher the temperature is in the open, the greater is the cooling effect of the forest. Under a forest stand during the summer, surface soil temperatures will be little, if any, higher than air temperatures, but in the open they may exceed air temperatures by as much as 50°–60° F. Data presented by Li (1926) showed that during July maximum surface soil temperatures under a 30-year-old eastern white pine stand in southern New Hampshire were 53.7° F. lower than in the open; during August, September, and October the differences were 42.0°, 30.4°, and 28.3° F., respectively. Wahlenberg (1930), working in Montana, reported maximum soil temperatures of 94° F. under *Ceanothus* brush and

145° F. in the open. In Arizona Shreve (1931) recorded soil temperatures of 105° F. in shaded situations and 124° F. in exposed situations, both measurements being taken at a depth of 3 in. In the Huron National Forest in Michigan maximum surface soil temperatures under partial shade during the summer are frequently 18°–30° F. less than in the open. Both summer and mean annual temperatures of soil are lowered by forest vegetation. The reduction probably amounts to about 5° F.

In considerations of the influence of forest vegetation on soil temperature it is necessary to recognize the effect of stand density. Summer temperatures under dense stands are lower than under open stands. As a rule, slightly higher temperatures are found in thinned stands than in unthinned stands, as Hoppe (1897), Ångström (1937), and Spaeth and Diebold (1938) have shown. The increase reported by Spaeth and Diebold as resulting from moderate improvement cuttings was slight and was confined to the humus layer. The increase reported by Ångström in heavily thinned stands of spruce in northern Sweden amounted to about 3.5°–5.5° F. Certain aspects of soil temperatures and stand density have already been discussed in Chapter 6.

Forest vegetation influences the development and persistence of frost in soils. Schubert (1930: 362) reported that freezing generally occurs earlier in bare soil than in soil supporting a forest. The data of other investigators confirm this view. Schubert also stated that frost does not penetrate as deeply in forest soils or remain as late as it does in soils which are bare. This situation prevails in some forests, but exceptions occur, particularly in regions of very low winter temperature and heavy snowfall. Belotelkin (1940) investigated the depth of freezing and the persistence of frost in soils of open fields and in several forest types in northern New Hampshire. He found the depth of freezing greatest in the spruce swamp type and least in the hardwood type. This fact may be explained by differences in the thickness of the snow mantle which accumulated under the stands. Dense stands of coniferous trees intercept considerable snow, whereas very little is intercepted by deciduous trees. Thawing of soil began and was completed earliest in the hardwood stand, followed in order by the open field, spruce flat, and spruce swamp.

The work of Li (1926) and others indicates that the influence

on soil temperature of a cover of grass is analogous to that of forest trees. However, a cover of moss or grass is usually less effective than trees are in reducing summer maximum temperatures at the soil surface.

NONLIVING COVER OF THE SOIL

The principal types of nonliving cover of soils are unincorporated organic debris and snow. By serving as insulating layers both tend to equalize soil temperatures.

It is quite certain that frequently a part of the influence on soil temperature ascribed to living forest vegetation is actually due to unincorporated organic matter. In an investigation in pure eastern white pine stands in southern New Hampshire Li (1926) reported that about 75 per cent of the reduction in maximum surface soil temperature was due to the forest canopy and about 25 per cent to the litter. On the other hand, he concluded that the effect of the forest in raising the minimum surface soil temperature was mainly due to the presence of the litter and that the forest canopy exerted little or no influence in this respect. MacKinney (1929) reported that litter in young Norway and eastern white pine stands reduced the day-to-day change in maximum surface soil temperatures by 50 per cent in the autumn and by 85 per cent in the spring. The influence of litter was to raise the mean maximum and mean minimum temperatures in the autumn and lower them in the spring. Litter reduced the depth of frost penetration by 40 per cent. Wood (1933) found that the mean maximum surface soil temperature in the oak-pine type of the New Jersey coastal plain was about 15.5°F . lower under a cover of oak leaves than it was where the litter had been removed. Measurements were made during the period July 5 to September 17.

The influence of a cover of snow on soil temperature is especially noteworthy. Snow, particularly if loose and fluffy, is a very good insulator because of its low thermal conductivity. Ramann (1911) gave the following heat conductivity values:

Fresh snow	0.00006
Old snow	0.0005
Water	0.0014
Ice	0.0057

The above values represent the amount of heat, in calories, transmitted per second through a layer having a thickness of 1 cm. and an area of 1 sq. cm. when the temperature difference is 1°C. Woeikof (1889) recognized that loose snow was a much better insulator than compact snow. The stabilizing effect on soil temperature of a blanket of snow, however, is probably not entirely due to insulating effects. As the temperature drops below 32°F., ice is formed, and the latent heat of fusion (79.9 cal. per gram) is liberated, which tends to return the temperature to 32°F. Likewise, when the air temperature rises and the snow melts, the soil is filled with water having a temperature of about 32°F. Largely because of the effects of snow, mean annual soil temperatures in cold climates are much higher than mean air temperatures.

Keen (1931: 328) reported that an air temperature of 5.0°F. at the surface of a 3-in. layer of snow may not produce any detectable change in the temperature in the soil. Beskow (1935) stated that a porous snow covering of about 18 in. is sufficient to prevent soil from freezing in northern Sweden; in southern Sweden freezing is prevented by a snow cover of about 8–12 in. A further illustration of the insulating effect of snow is found in northern Michigan.⁵ The water pipes between buildings at the Upper Peninsula branch of the Lake States Forest Experiment Station are placed at an average depth of 14–16 in. in the soil. They never have frozen, although minimum air temperatures of more than -40°F. have been recorded. The average depth of snow during January, February, and March is about 18–24 in.

Mean monthly soil temperatures in the United States are higher during the winter at northerly stations, where the snow cover is thick, than at stations further south or west, where the snow cover is thin or lacking (Fitton and Brooks, 1931). Ångström (1937) reported a deeper snow cover in thinned stands of spruce in northern Sweden than in unthinned stands. Because of the protection afforded by the thick cover of snow, soil temperatures in the thinned stands were higher than in dense stands where the snow cover was less. Frozen soil in the thinned stands thawed 2–4 weeks earlier than in unthinned stands. In the spring, thawing of the soil proceeds from above and from

⁵ Personal communication from John R. Neetzel, Lake States Forest Experiment Station.

below. Thawing at the surface does not begin until the snow has melted; thawing from below may begin earlier.

INTERNAL FACTORS INFLUENCING SOIL TEMPERATURE

The temperature of a soil body is influenced by certain characteristics of the soil itself. Chief among them are specific heat, conductivity of heat, color, and content of water.

SPECIFIC HEAT

Specific heat refers to the calories of heat necessary to raise a unit weight (grams) or volume (cubic centimeters) of a substance 1°C . From the standpoint of the ecologist specific heat values on the basis of volume are more significant than values on the basis of weight. The volume, not the weight, of soil heated is important. On a weight basis humus has a higher specific heat than mineral soil, but on a volume basis the situation is reversed, as the following data from Ulrich (1894) indicate:

Material	Specific Heat, weight basis	Specific Heat, volume basis
Sand	0.191	0.292
Clay	0.224	0.233
Humus	0.443	0.165

The specific heat of soil in the field is profoundly influenced by the amount of water present. As the water content increases, more heat is required to cause a rise in temperature of 1°C . This phenomenon is explained by the fact that the specific heat of water itself is very high, being about four or five times that of most dry soil material.

CONDUCTIVITY OF HEAT

It appears that the thermal conductivity of dry mineral soil material varies but little from one soil to another (Smith and Byers, 1939). Factors such as the content of organic matter, the texture, volume weight, and porosity are more important than the composition of the mineral material. A high content of organic matter indicates low thermal conductivity. Soils hav-

ing a sandy texture tend to have higher conductivity than do clays. As volume weight increases and porosity decreases, there is an increase in the transfer of heat.

It is well known that the thermal conductivity of dry soil is increased when water is added. The reasons for this are: (1) the thermal contact is improved since the soil particles are held more closely together by surface tension, and (2) although water is not as good a thermal conductor as are soil particles, it is a much better transmitter of heat than is air. With water present at the points of contact between soil particles, heat can move through the water by conduction more readily than through soil air by convection or radiation. Patten (1909) stated that heat will pass from a soil grain to soil water 150 times easier than from a soil grain to the soil air. As the water content of a soil increases, the conductivity will gradually approach the value for water itself.

Dry soils have relatively low specific heat, but their thermal conductivity is also low; consequently the temperature drops off rapidly with increasing depth below the heated surface. On the other hand, moist soils have relatively high thermal conductivity, but because of their high specific heat they do not attain high temperatures.

The thermal conductivity of rocks is higher than that of dry soil material. This fact, as already pointed out, means that a soil containing rocks heats up more during the summer than does a soil having few, if any, rocks. On the other hand, rocky soils tend to cool rapidly during periods of cold weather. The result is that temperature fluctuations in a soil body tend to increase as the content of rock increases. Larsen (1930) stated that soils occupied by the ponderosa pine type in the northern Rocky Mountains invariably contain much rock near the surface. He expressed the view that the rocks favored high soil temperatures during the summer and that the soil temperature fluctuations were greater than in other forest types.

COLOR

Soil color has an effect on both heat absorption and heat radiation. In general, dark-colored soils are better absorbers and better emitters than are light-colored soils. Consequently, tem-

peratures of dark-colored soils are usually higher than those of light-colored soils. Isaac (1930) measured temperatures of yellow mineral soil and black charcoal surfaces in the Wind River Valley in Washington. His data, which follow, indicate that the charcoal surfaces attained much higher temperatures than the yellow soils.

Date (1929)	Black Surface Temperature, °F.	Yellow Surface Temperature, °F.
May 20	143	125
May 21	144	135
May 22	144	137
May 23	134	122
May 24	133	121

The influence of color on temperature is greatest when soils are dry; with increasing content of moisture the influence of color decreases. Dark-colored soils with a high moisture content are usually cooler than light-colored soils which are dry.

CONTENT OF WATER

The temperature relations of a soil are profoundly influenced by its moisture content. It has already been pointed out that the specific heat and thermal conductivity of dry soil are increased when water is added.

Certain other properties of water which have a bearing on temperature relations need to be recalled. Changes from the solid to the liquid state, the solid to the vapor state, and the liquid to the vapor state all involve absorption of heat. This heat is referred to as latent heat of fusion, latent heat of sublimation, and latent heat of vaporization, respectively. Reverse changes, that is, liquid to solid, vapor to solid, and vapor to liquid, involve liberation of heat. At 0°C. the latent heat, in gram-calories, is as follows: fusion, 79.7; sublimation, 675.6; vaporization, 595.9. The effect of water in soil is to greatly reduce temperature fluctuations.

Moist or wet soils are usually cooler than dry soils. The widespread recognition of this fact is indicated by the layman's characterization of relatively dry sands as "warm" soils and relatively wet clays as "cold" soils. Belotelkin (1940) found

that frost penetrates more deeply and remains longer in poorly drained soils than in those having good drainage.

During the summer maximum temperatures at or near the surface of soils with a water-table at shallow depth are usually lower than in soils having a water-table at greater depth (Feilberg, 1937; Stoeckeler and Sump, 1940). Stoeckeler and Sump invariably found 2-4 per cent more available moisture in the upper 6 in. of soils having a water-table within 2-5 ft. of the surface than in soils having a water-table at a depth of more than 6 ft. They also found that maximum surface temperatures were much less in soils having a water-table near the surface than in soils where the water-table occurred at considerable depth. In the former soils maxima of 100°-120°F. were recorded, but in the latter maxima reached 140°-145°F. Differences of this magnitude are highly important from the standpoint of seedling survival. In the Lake States region it has been found that day-time watering in nurseries results in an appreciable reduction of surface soil temperatures.

DIURNAL VARIATIONS IN SOIL TEMPERATURE

As a general rule, the upper part of a soil body experiences a daily rise and fall of temperature. This diurnal variation in temperature is greatest in the summer and least in the winter. In fact, during the winter little, if any, diurnal change can be noted.

During the day the surface soil heats up, and a difference in temperature between the surface and deeper layers is thus established. A heat wave is propagated into the soil body. At night, when the soil surface cools below the temperature of the deeper layers, heat moves upward. The movement of heat is from zones of relatively high to zones of relatively low temperature. Heat waves travel slowly and are considerably "damped" in their downward passage; with increasing depth below the surface the amplitude of the temperature range is diminished. Diurnal variations are noted to a depth of about 3 ft.

As a result of the diurnal movement of heat the temperature at any given soil depth exhibits fluctuations which may be regarded as a reduced and delayed image of the temperature fluctuations at the surface. Minor fluctuations in temperature

at the surface will not be reflected in the lower horizons. Although the amplitude of the surface variations diminishes with increasing depth, and there is a time lag in the appearance of the temperature change in lower horizons, the time period of the wave remains practically unaltered. Li (1926) found that the delay in response of temperature in the soil body to changes on the surface usually amounted to 1 hour per inch of soil depth through which the heat passed. McCulloch and Hayes (1923) reported that at Manhattan, Kansas, the response at the 1-ft. depth usually occurred on the same day; at greater depths the response was approximately 1 day later for each foot of soil.

Diurnal changes in soil temperature are greatly influenced by soil cover; variations are greatest in bare soils. Clear, hot days and clear, cool nights favor wide temperature fluctuations. In the summer during the day maximum surface soil temperatures commonly rise far above maximum air temperatures, but at night minimum soil temperatures do not fall as low as minimum air temperatures.

SEASONAL VARIATIONS IN SOIL TEMPERATURE

An analogy may be drawn between seasonal fluctuations in temperature and diurnal fluctuations. During the spring and summer the soil surface heats as it does during the day; in the autumn and winter the soil surface cools as it does during the night.

As a general rule, the greater is the seasonal amplitude of surface soil temperature change, the greater is the depth at which fluctuations are noticeable (Lang, 1926). Fitton and Brooks (1931) reported a seasonal amplitude of 10°-12°F. at a depth of 10 ft. near Bozeman, Montana. Smith (1932) found the following seasonal ranges in the temperature of a bare soil near Davis, California:

Depth of Soil, feet	Range in Temperature, °F.
4	27
5	22
6	18
8	14
10	12
12	9

The depth to which seasonal variations are noticeable varies greatly. Lang (1926) reported fluctuations in temperature at a depth of about 25–40 ft. in Germany, and Ijjász (1938) stated that variations were noticeable at depths of about 65–82 ft. in Hungary.

FIRES AND SOIL TEMPERATURE

The influence of fires on surface and subsurface soil temperatures varies greatly. This situation is to be expected in view of the wide variations in duration and intensity of fires and differences in the physical nature and moisture content of mineral soils.

Elpatievsky *et al.* (1934) measured soil temperatures developed as a result of burning piles of slash in pine and spruce forests in Russia. The following temperatures were recorded when a pile of spruce slash about 5 ft. in height and 7 ft. in diameter on a heavy-textured soil was burned.

Temperature at the surface under the center of the pile, 1382°F.

Temperature at the surface near the periphery of the pile, 1274°F.

Temperature at a depth of $\frac{3}{4}$ in., fluctuated from 77° to 217°F.

Temperature at a depth of $1\frac{1}{2}$ in., fluctuated from 64° to 185°F.

Temperature at a depth of $7\frac{3}{4}$ in., fluctuated from 54° to 86°F. The horizons attaining the highest temperatures were the litter layers and the zone between the A_0 and A_1 . Isaac and Hopkins (1937) investigated temperatures attained during broadcast burning of Douglas fir slash in the Wind River Experimental Forest, near Carson, Washington. The highest temperatures recorded immediately above and 1 in. below the forest floor (1.5 in. thick) were 1841° and 608°F., respectively. These temperatures were recorded at one of the points where the slash was heaviest.

In an investigation of temperatures during forest fires in the longleaf pine region Heyward (1938) found that temperatures as high as 200°F. were infrequent at a depth of $\frac{1}{8}$ – $\frac{1}{4}$ in. in the mineral soil. Temperatures at these depths usually ranged from 150° to 175°F., and the maximum temperature persisted for only 2–4 minutes. The maximum temperature recorded at a depth of $\frac{1}{2}$ in. was 195°F. Heyward concluded that the majority

of fires in the longleaf pine region do not heat the mineral soil sufficiently to cause impoverishment.

Indications are that forest fires seldom result in temperatures of more than about 175°-200°F. at depths greater than 1 or 2 in. in the mineral soil. Exceptions to this generalization may occur in situations where the mineral soil is dry and where large amounts of slash are concentrated.

ECOLOGICAL SIGNIFICANCE OF SOIL TEMPERATURE

Soil temperature influences the germination of seeds and the survival and development of seedlings. Even if moisture and aeration conditions are favorable, seeds will not germinate if the temperature is too low. Adams (1934) reported a germination period of 11 days for eastern white pine in soil having a mean surface temperature of 88°F.; in soil having a temperature of 57°F. the period was 63 days. It was also found that root growth of year-old white pine seedlings was greater and the top-root ratio smaller when the soil temperature was comparatively high (above a mean temperature of 55°F. at an 8-in. depth, increasing to a mean surface temperature of 88°F.) than when these conditions did not hold. It is probable that soil temperatures under dense forest stands are frequently below the region of the optimum for germination of seeds and development of seedlings. Poor development of reproduction in shade may be partly explained by low soil temperature (Bühler, 1918: 377).

In situations exposed to full, or nearly full, solar radiation, surface soil temperatures frequently rise high enough to be fatal to young seedlings. The temperature at which death occurs varies with the species and the condition of the seedlings. Münch (1913) concluded that temperatures exceeding about 129°F. commonly were lethal. Toumey and Neethling (1924) stated that loss of seedlings may occur when the surface temperature exceeds 122°F. Seedlings of representative conifers of western America were quickly killed when a temperature of about 130°F. was reached (Baker, 1929). Temperatures greatly exceeding these values frequently occur in nature and are known to result in seedling mortality.

Temperature exerts indirect, as well as direct, effects. At low temperatures water uptake is inhibited, and at high tempera-

tures water losses by evaporation from the soil may be excessive. Both the fauna and flora of the soil are influenced by soil temperature, as has been pointed out in Chapter 5.

MEASUREMENT OF SOIL TEMPERATURE

A wide variety of instruments are available for measuring soil temperatures. The most simple are mercurial or alcohol thermometers, by means of which current, maximum, and minimum temperatures may be determined. Thermographs of many types have been used in recent years. These instruments are extremely useful because they permit evaluation of the duration and intensity features of temperature. The use of resistance thermometers has been discussed by Harrington (1928) and Smith (1932). Mail (1933, 1935, 1937) and Spaeth and Diebold (1938) found that copper-constantan thermocouples gave satisfactory results in soil temperature investigations.

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Chapter 9

THE WATER RELATIONS OF SOILS, PARTICULARLY FOREST SOILS

Water relations occupy a position of outstanding importance because of their influence on the various physical, chemical, and biological properties of soils. Water serves as a solvent and medium of transport for plant nutrients and supplies an essential need of all living organisms. Even the richest of soils, as judged by mineral nutrient content, fail to support forests unless the minimum water requirements are satisfied. Conversely, few, if any, soils are so poor in nutrient elements that they are unable to support forest vegetation when an adequate supply of water is present.

The primary factors determining soil-water relations are climate, topography, nature of the soil, and vegetation. Their influence will be considered in this chapter. By referring to Figure 33, the student can gain a general view of the principal features of the water cycle.

SOME GENERAL CONCEPTS OF SOIL MOISTURE

All soils contain pores of varying size and configuration. These pores may consist largely of the interstices between the solid soil grains, as in very sandy soils, or they may consist of the spaces between and within aggregates of soil particles. Relatively large pores, representing passageways of soil animals or old root canals, are particularly common in forest soils and have an important bearing on water relations. In all soils the pores form an exceedingly complex system which is normally occupied partly by air and partly by water.

Two forces are responsible for the retention of water in soils: adhesion and cohesion. Adhesion involves the attractive force of molecules in the surfaces of the soil particles for water molecules, and cohesion concerns the attraction of water molecules for other water molecules. Through the action of the forces of

adhesion very thin, tightly held films of water are built up on the surfaces of the soil particles. As more water molecules are attracted by the forces of cohesion, these films increase in thickness and finally fill the capillary pores unless they are prevented by entrapped air. If free drainage is provided, the large non-

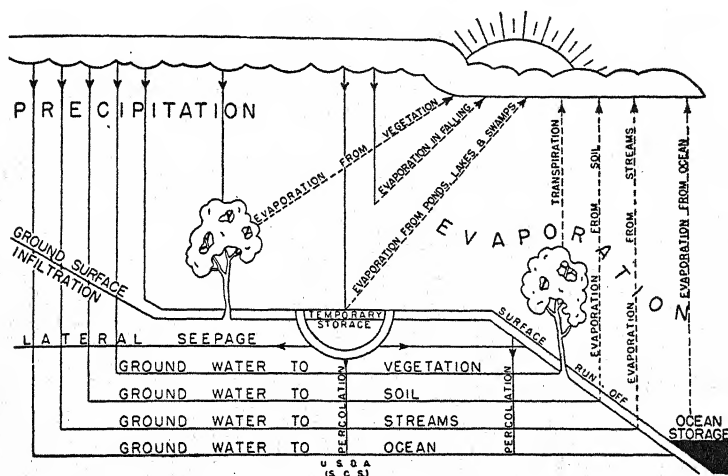


FIG. 33. The hydrologic cycle. Courtesy, United States Soil Conservation Service.

capillary pores are not completely filled with water because of the pull of gravity, but they become lined with thick films. The attractive force of water molecules for each other is clearly manifested at liquid-air interfaces by the phenomenon known as *surface tension*. Its importance will be discussed later in connection with the movement of soil water.

The degree of tenacity with which soil moisture is held by adhesion and cohesion varies, depending on the position of the water. That adsorbed on the soil particle surfaces is so firmly bound that it is in a nonliquid condition and unavailable to plants. Proceeding outward from the soil-water interface, the water molecules are held by progressively weaker forces until, at the water-air interface of a thick film, surface tension alone constitutes the retaining force. Thus relatively little force is required to remove a unit volume of water from a soil having a moisture content equivalent to its field capacity; but, as the moisture content de-

creases, the force needed to remove unit volumes of water becomes ever greater.

The energy relations of soil water have been a subject of inquiry by soil scientists for many years. Interest in the matter has been based on a general recognition of the importance of these relations in the movement of soil water and in determining its availability to plants. With the introduction by Schofield (1935) of the pF scale for specifying the energy with which water is held the subject has gained new interest and concepts have been simplified.

THE pF SCALE

From what has been said it will be apparent that all soil moisture is not held with equal tenacity. It will also be understood that the force required to remove a given increment of water from a soil will represent the force by which that water was held. This force, for any soil moisture content, may be expressed as the height in centimeters of a water column having an equivalent weight.¹ Practical difficulties are encountered, however, when energy values are expressed directly in the above terms. For example, the height of a water column equivalent to the force with which moisture is held at the surfaces of colloid particles by adhesion would exceed 1,000,000 cm. Values of this magnitude are so cumbersome that it is customary to use instead the logarithm of the equivalent column. These logarithmic expressions are called pF values (Schofield, 1935). The symbol p in the expression indicates the logarithmic nature of the value; the symbol F represents free energy or force. Designation of the energy relations of soil water in terms of pF is analogous to specification of reaction in terms of pH .

Low pF values correspond to low energy values, and, conversely, high pF values signify high energy values. Thus a pF of 2 indicates that the force with which water is retained is equivalent to the weight of a water column 100 cm. in height, a pF of 3 to a column 1000 cm. in height, etc., on to pF 7.0, which corresponds to a column 10,000,000 cm. in height. It should be recognized that a given pF value indicates the energy relation at

¹ Buckingham (1907) used the height of an equivalent water column as an expression of what he designated the capillary potential, a concept which is considered in the section dealing with the movement of soil water.

the water-air interfaces and does not apply to the soil water as a whole.

The relation between pF values, height of an equivalent water column in centimeters, and atmospheres of pressure is given in Table 35 from Lyon and Buckman (1943: 150). Several decimal values for pF are recorded because they correspond to certain so-called soil moisture constants, which will be discussed later.

TABLE 35. A TABLE OF pF EQUIVALENTS*

(After Lyon and Buckman, 1943:150)

pF Values	Height of a Unit Column of Water, centimeters	Approximate Atmospheres of Pressure
0	1	$\frac{1}{1000}$
1	10	$\frac{1}{100}$
2	100	$\frac{1}{10}$
2.7	501	$\frac{1}{2}$
3	1,000	1
4	10,000	10
4.2	15,849	15
4.5	31,623	31
5	100,000	100
6	1,000,000	1,000
7	10,000,000	10,000

* From *The nature and properties of soils*, by T. L. Lyon and H. O. Buckman. By permission of The Macmillan Company, publishers.

THE SIGNIFICANCE OF pF CURVES

Typical pF curves for several soils of different texture are presented in Figure 34. The curves in Figure 34 show that the amount of water held at a given pF differs greatly in soils of different texture. For example, at a pF of 4.2, which represents the maximum tension at which plants can obtain water, clay contains approximately 19 per cent moisture, silt loam about 10.5 per cent, fine sand about 4 per cent, and quartz sand about 1 per cent. These values are, of course, only approximate; permissible variations of particle size within a soil texture class and other features, such as organic matter content and structure, will affect the shape of pF curves.

The general shape of a pF curve is determined by soil texture, as a result of its influence on pore-size distribution. In the

process of removing water from a soil by pressure or suction, the large pores are emptied first and the smaller ones later. The flat portion of the curve in Figure 34 for the quartz sand separate is explained by the fact that most of the pores in this material were of uniform size. When tension sufficient to empty this particular size of pore was applied, most of the soil water moved out. In the example under consideration this occurred at about pF 1.7. The relatively small amount of water removed from the quartz sand at pF values of less than 1.7 came from pores of

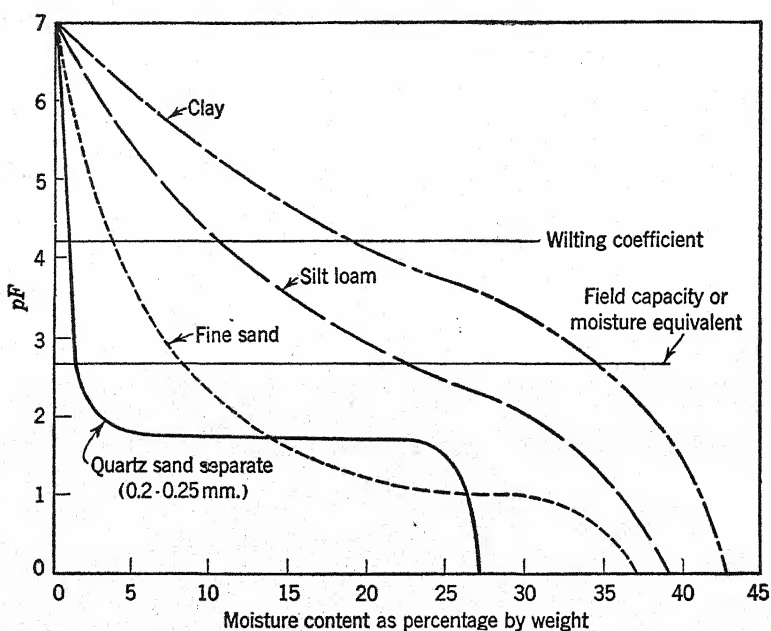


FIG. 34. pF curves for several soils differing in texture. Data for the quartz sand separate and the clay and silt loam soils are from Woodruff (1941); the curve for the fine sand soil is reprinted by permission from *Soil physics*, by L. D. Baver, 1940. John Wiley and Sons, Inc.

larger than average size, and that removed at pF values greater than 1.7 came from pores of smaller than average size.

All the pF curves presented in Figure 34 are drying curves; that is, the soils were initially saturated and then dried by removing water to obtain the pF values. If the soils had been

dry at the start and then wetted, the pF values, and consequently the curves, would have been somewhat different, although the general relationships would be similar. At the same moisture content the pF is greater in a soil which is being dried than in one which is being wetted. To express the situation differently, it may be said that for a given moisture content of a soil there are two pF values, one based on drying and the other on wetting. Only values based on drying are considered here.

A pF curve is extremely helpful in evaluating the moisture relations of a soil. By inspection the approximate amount of water the soil will contain when saturated (pF 0), when at field capacity (pF $2.7 \pm$), and when plants will wilt (pF 4.2) may be ascertained. Various methods have been developed for measuring pF values in soils. For details of the different techniques the work of Schofield (1935), Baver (1940: 205-207), Woodruff (1941), and Richards and Weaver (1944) should be consulted.

THE CLASSIFICATION OF SOIL WATER

The pF curve represents a continuous variation in tension as the soil changes from a moist to a dry condition. In other words, clear-cut boundaries which might serve to distinguish moisture conditions are lacking. Nevertheless, it is convenient to recognize various categories of soil water which fall within certain broad energy ranges. This is made evident by the fact that several classifications of soil water have been employed by soil investigators (Briggs, 1897; Bouyoucos, 1917, 1918, 1921; Parker, 1922; Lebedeff, 1928; Keen, 1931; Baver, 1940). One of the earliest classifications, and still the most widely used, is that of Briggs, who recognized three forms of water:

Hygroscopic water: that portion which is retained as a thin film by the soil particles after capillary water has been removed; it is incapable of movement by gravitational or surface tension forces.

Capillary water: that portion which is retained around the soil particles and in the capillary pores after gravitational water has moved out; it is capable of movement under the influence of surface tension.

Gravitational water: that portion which the soil is incapable of retaining under conditions of free drainage; it drains away under the influence of gravity.

Of these three forms of water the second is of greatest ecological importance. In soils having good drainage gravitational water is transitory and not very important to plants. Hygroscopic water is held in the soil by forces so great that plants are unable to remove it.

It should be borne in mind that the divisions drawn between different forms of soil moisture in the foregoing classification are not precise. There are no sharp boundaries between the different forms; each grades by insensible degrees into another. The different "forms" of soil water outlined serve as convenient designations for parts of the moisture range and as such seem justifiable.

HYGROSCOPIC WATER

Most of the hygroscopic water in a soil is associated with the organic and inorganic colloidal material. It is frequently visualized as existing in the form of a thin film of not more than 15-20 layers of water molecules on the soil particles. Ordinarily the greater is the content of colloidal material in a soil, the higher is the content of hygroscopic water (Keen and Coutts, 1928; Coutts, 1929). It should be recognized, however, that both the chemical and the physical nature of the colloids are influential in determining hygroscopicity. Consideration of this fact has tended to discourage use of hygroscopicity values in estimating the quantity of colloid present and the surface area involved. The hygroscopic moisture in fine clays may amount to 20 per cent or more; in sands it may be less than 1 per cent. Hygroscopic water may be conveniently considered as that water held by the soil at forces greater than a pF of 4.5.

CAPILLARY WATER

Capillary water occupies the small pores in a soil. In this water, which is in intimate contact with the solid particles of the soil, occur important chemical and biochemical transformations. The water requirements of forest trees are supplied, for the most part, by capillary water. Capillary pore space is particularly important from the standpoint of the hydrologist, for it represents the retention storage capacity of the soil. Water held in the capillary pores does not directly contribute to streamflow.

Various factors influence the amount of capillary water which a soil can hold. With increasing fineness of texture there generally is an increase in the amount of capillary pore space. Clays hold far greater amounts of capillary water than do sands. The capillary capacity of a soil is also influenced by structure. A compacted sandy soil holds more capillary water than a loose sandy soil. On the other hand, a compacted clay holds less than one with well-developed aggregates. Organic matter, because of its highly porous nature, has a pronounced tendency to increase the capillary capacity of coarse-textured soils. Henry (1908: 59) reported that organic matter from a spruce forest absorbed over four times its own weight of water. Keen and Coutts (1928) found that organic colloidal material took up 4.4 times its own weight of water, and inorganic colloidal material took up 2.7 times its weight. Although these values must be regarded as very general, they indicate that organic matter is capable of absorbing very large amounts of water.

The tendency of organic and inorganic colloids to swell on wetting results in a reduction in the diameters of the soil pores. This decrease favors greater capillary capacity.

Soil depth also influences the amount of capillary water which may be held (Keen, 1931: 204). The deeper is the soil, the greater are the curvatures which water films in the upper part of the body have to assume to balance the downward pull. Consequently, less capillary water is held in the upper layers of a deep soil body than in one more shallow. Capillary water is held at forces between a pF of approximately 2.7 and 4.5, although at the higher pF values capillary movement is very sluggish and of little use to plants.

GRAVITATIONAL WATER

The addition of water to a soil body whose capillary capacity is already satisfied results in the appearance of gravitational water. This water, under the influence of gravity, moves down through the noncapillary pores, its movement being governed by the transmission rate of the soil. The noncapillary porosity of a soil is of hydrologic significance, because the large pores provide for detention storage of water. Water in detention storage, unlike that in retention storage, is only temporarily de-

tained; it moves out under the force of gravity as opportunity for drainage occurs.

If a limited amount of water is added to the surface of a dry soil having uniform texture and structure, it does not moisten the entire soil body; instead, the moisture content of an upper layer is raised to the neighborhood of its capillary capacity, and below this layer little effect is noted (Veihmeyer and Hendrickson, 1934). The thickness of the upper layer which is moistened depends on the capillary capacity of the soil and the amount of water added. With the further addition of water the lower boundary of the wetted zone moves downward. Finally, when the wetted zone extends deeply enough, the excess moisture passes into the ground water. It does not follow that gravitational water moves out of a forest soil only after the capillary capacity of the upper layers has been satisfied. Most forest soils are not structurally uniform; they have large noncapillary pores which serve admirably as channels for the rapid movement of water, with the result that after precipitation the lower boundary of the wetted zone is extremely irregular. Water may thus drain down to the water-table through a soil body which, in parts, is still dry. Deep-rooted tree species favor vertical drainage in heavy-textured soils (Müller *et al.*, 1936: 730). The canals left after roots decay serve as drains for the removal of water.

Drainage of gravitational water may be inhibited by reason of topographic situation or the presence of impermeable strata in the soil body. This situation commonly results in a high water-table.

The general relationships between gravitational, capillary, and hygroscopic water are presented in Figure 35, which is taken from Lyon and Buckman (1943: 155). Note that gravitational water is held by forces somewhat less than a p^F of 2.7.

WATER-TABLE RELATIONS

At variable depths below the soil surface a zone may be encountered in which the soil or underlying permeable material is saturated with water under hydrostatic pressure. This is the *zone of saturation*, and the contained water is called *ground water*. The *water-table* is the upper surface of a zone of saturation

(Meinzer, 1923). Ground water held in a position above the main zone of saturation is said to be perched, and its water-table is designated as a perched water-table. Perched water-tables result when the downward movement of gravitational water to the main water-table is prevented by impervious strata. Extending upward from the water-table is a belt known as the *capillary*

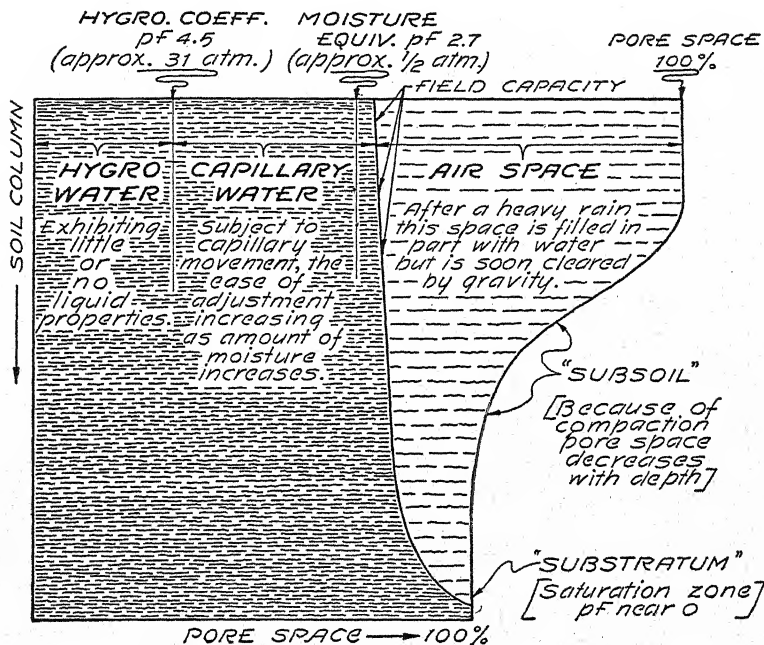


FIG. 35. Diagrammatic representation of the general relationships between gravitational, capillary, and hygroscopic water in a soil column of uniform texture well drained at the top but saturated at the bottom. Gravitational water, when present, occupies the air space at the right. From *The nature and properties of soils*, by T. L. Lyon and H. O. Buckman. By permission of The Macmillan Company.

fringe. In this belt the small pores are filled with water that is continuous with the water in the zone of saturation. Water in the capillary fringe is held by capillarity acting against the force of gravity. The thickness of the capillary fringe varies, depending on the texture of the soil material; in coarse materials it is very thin, and in fine materials it is relatively thick.

The profile of the main water-table bears a resemblance to the profile of the ground surface, particularly in regions of gentle topography. The more accentuated is the topography, the greater is the deviation between the profile of the land surface and the profile of the water-table. Likewise, when openings in the soil or rock material are large and numerous, there are greater differences between the land-surface profile and the water-table profile than when the openings are small and few. In areas which are not much higher than adjacent bodies of water, the water-table is near or at the surface; in elevated situations it may lie so deeply that no influence on vegetation is apparent.

Ground water is drawn upon by trees if the water-table or the capillary fringe is within reach of their roots. In view of this fact it is not surprising that the water-table has been observed to stand lower under forest stands than in adjacent open areas (Ebermayer, 1889).

It might be expected that the removal of forest stands, by clear-cutting or other means, would result in a rise of the water-table. This rise occurs in situations where the ground water is close enough to the surface to be reached by tree roots. Swamp conditions may be produced in areas where the water-table, even before forest removal, was close to the surface (Bühler, 1918: 321; Wiedemann, 1925; Gulisashvili and Stratonovitch, 1935). When swamp conditions prevail, regeneration of desirable species becomes very difficult, and productivity of the site is lowered. In areas susceptible to swamping, complete removal of forest stands should be avoided.

The position of the water-table fluctuates in the course of time. Fluctuations are generally due to variations in transpiration, evaporation, precipitation, atmospheric pressure, and temperature. In the temperate zone a rise during the winter and a drop during the summer are generally observed (Lang, 1926: 395). Diurnal changes of the water-table level have been recorded in forested swamps in New Hampshire. A rather consistent daytime drop and nighttime rise were found to occur. Presumably changes of this type are associated with transpiration.

Water-table position may be permanently raised or lowered as a result of human influences. The construction of dams raises the ground water in adjacent areas, and drainage or use for domestic and industrial purposes lowers it. Lang (1926: 291)

reported decreased forest production in parts of Europe as a result of the utilization of ground water for domestic and industrial purposes. Krauss (1928) mentioned the removal of ground water by mine tunnels and other means as possibly contributing to the difficulties of spruce culture in Saxony. The water-table may also be lowered as a result of gully erosion.

MOVEMENT OF SOIL WATER

Water moves through the soil pores in both liquid and vapor form. The agents responsible for the movement of water in the liquid phase are gravity and capillary tension. Vapor pressure differences are responsible for movement of water in the vapor phase.

MOVEMENT OF THE LIQUID PHASE

The force of gravity causes free water in the noncapillary pores to move downward in a soil body until it reaches an impermeable layer or the water-table. Two factors are of importance in this connection: (1) the number, size, and continuity of the noncapillary pores, and (2) the extent to which air is entrapped in the pores by the entering water.

With greater number, size, and continuity of the noncapillary pores percolation increases. Measurements repeatedly have shown that there is a relation between the volume of noncapillary pores and infiltration rates. It may be pointed out that the horizons in a soil frequently differ in noncapillary porosity and that the rate of percolation is limited by the horizon which is least permeable. If air is entrapped in the soil pores by the entering water, the rate of percolation is reduced. This is most apt to occur when very heavy rains follow periods of drought.

Relatively free movement of gravitational water in the soil body is desirable from the standpoint of water infiltration and internal drainage. The higher is the proportion of precipitation which enters the soil, the lower will be the surface run-off and erosion. Good internal drainage is necessary if a soil is to be well aerated.

As a rule, lateral movement of ground water under the influence of gravity is relatively slow. Movement is more rapid on slopes

than in situations where the land is flat. Likewise, movement is more rapid in highly porous strata, such as gravel and sand, than in fine material, such as clay. The movement of ground water is important from an ecological point of view. Higher concentrations of oxygen are usually found in ground water which is moving than in that which is stagnant. An excellent discussion of the theory of ground-water motion may be found in the paper of Hubbert (1940).

The movement of water under the influence of capillary tension takes place in the capillary pores of a soil. In contrast to the predominantly downward movement of water which is being acted upon by gravity, movement due to capillary tension may occur in any direction, vertically or laterally.

Two types of interfaces are found in moist soils. One type is formed by water in contact with the soil particles; this is the soil-water interface. The second type is formed by water in contact with the soil air; this is the water-air interface. Water molecules at the soil-water interface are subject to the strong force of attraction called adhesion. Water molecules at the water-air interface are subject to differential forces inward, due to molecular attraction or cohesion. The effect produced, called surface tension, is similar to that of a stretched elastic membrane over the liquid surface, acting to reduce the surface area to a minimum. In capillary pores the curvature of the water-air interfaces is convex toward the water. This fact means that there is a negative pressure or pull on the water in the capillary pores. The greater is the curvature of the water-air interface, the greater will be the pull exerted.

Capillary rise of water is greatest in soils having small capillary pores. Thus water will rise higher above the ground-water level in clays than in sands. On the other hand, the initial rate of movement is greater in sands than in clays.

In 1897 Briggs developed the idea that capillary water occurred as a thin, continuous film around the soil particles. At the points of contact between particles the curvature of the films, or water-air interfaces, is concave outwards. The movement of water from zones of high water content to those of low content was related to differences in the thickness of the films or, in other words, to differences in the curvature of the films.

With the work of Haines (1927, 1930) emphasis was shifted

from the configuration of water films around soil particles to the configuration of water films in the soil pores. Haines viewed the soil pores not as thread-like capillary tubes but as a complex assemblage of cellular units of varying size which communicate with each other through narrow necks. It was found that changes in moisture content at a given point in a soil do not proceed smoothly; rather, movement of water in or out of the irregularly shaped pores shows two unstable stages, at which filling or emptying is completed suddenly. This is explained by the fact that a water-air interface in a narrow neck between two cells has greater capillary pull than an interface in the cells themselves. Thus, when water entering a cell has passed the widest part, the inflow may increase abruptly, with a resultant sudden completion of filling.

A further important contribution of Haines was the concept that changes in moisture content are for the most part not strictly reversible but fall into two series, corresponding to two directions of moisture change. In general, a soil offers a greater capillary pull against the extraction of water from its pores than is developed when water is being drawn in. In other words, the suction required to remove water from a moist soil is greater than the force against which water enters. The suction exerted by dry soil is only sufficient to remove water from a considerably wetter soil layer. Pressure deficiencies at a given moisture content may differ materially, depending on whether the soil is in the rising or falling moisture stage. A pronounced hysteresis is noted in the wetting and drying curves.

The attraction of soil for water has been designated as the capillary potential by Buckingham (1907). Other writers have referred to it as suction force or suction pressure. Buckingham expressed the capillary potential in terms of the equivalent water column which corresponded to the pressure deficiency necessary to abstract water from a soil, or to the positive pressure necessary to move water into a soil at a given moisture content. Capillary potential increases as moisture content of a soil decreases; water tends to move from zones of low to zones of high capillary potential. When heating occurs at any point, the surface tension and viscosity of the water are lowered, and it may be drawn to regions of higher surface tension. In terms of pF this statement means that water tends to move from regions of low pF to those

of a higher pF , since pF is simply the logarithm of the capillary potential.

In the earlier literature great importance was attached to capillarity as a mechanism for the distribution of water in soils. Particular emphasis was frequently placed on the role of capillarity in raising water from a water-table or moist subsoil. It now appears that the importance of capillarity in moving water through considerable distances was exaggerated. Keen (1928) carried out a careful investigation of the extent of capillary rise in soil and concluded that for all practical purposes upward capillary movement is ineffective over a distance greater than about 35 cm. in coarse sand, 70 cm. in fine sand, and 85 cm. in heavy loam. It was suggested that, in evaluating the total depth of soil supplying moisture for plants, about 3 ft. should be added to the average depth of penetration by roots.

In general, capillary movement of water from moist soil to dry soil is too slow to be of much significance for plants (Alway, McDole, and Trumbull, 1919; Veihmeyer, 1927; Veihmeyer and Hendrickson, 1934). If plants are to utilize the moisture in a given body of soil, they must extend their roots into that soil. It is likely, however, that movement from the zone of saturation into the belt known as the capillary fringe is fairly rapid.

The ineffectiveness of capillarity as a mechanism for the movement of water from moist to dry soil needs to be considered by investigators who attempt to maintain soils in pot cultures at moisture contents below the field capacity. Maintenance of such moisture conditions seems impossible of attainment.

MOVEMENT OF THE VAPOR PHASE

Water in the vapor phase may move from one point to another within the soil body. It may also move from the atmosphere into the soil or from the soil into the atmosphere; movements of these two types are considered in a subsequent section.

Movement of water vapor is due to vapor pressure differences and proceeds from points of high to points of low pressure. Differences in vapor pressure arise as a result of differences in moisture content or temperature.

As long as the soil moisture content remains above the hygroscopic coefficient, the relative humidity of soil air is main-

tained at or near 100 per cent (Lebedeff, 1928). Consequently, movement of water vapor in the soil, due to variations in moisture content, would be expected only when the soil moisture content was below the hygroscopic coefficient (pF 4.5 or more). Movements of water vapor resulting from variations in soil moisture content are believed to be relatively unimportant in forest soils.

The influence of differences in temperature on vapor pressure, and consequently on movement of water vapor in soils, may, however, be significant. Smith (1940, 1944) found that there was considerable movement of water in the vapor phase when small thermal gradients were applied to soils having moisture contents in the region around 99 per cent relative humidity and extending to above 50 per cent of the moisture equivalent. During periods when there is strong nocturnal cooling of the surface soil layers, the vapor pressure is reduced to a point below that in the warmer, deeper-lying horizons. Under these conditions water vapor may move upward and be condensed. If the surface soil layers are warmer than the subsurface layers, downward movement may occur. This phenomenon is probably more important in bare soils or those supporting grass than in those supporting forest vegetation; temperature changes in the latter are relatively small. Lebedeff (1928) attached considerable importance to the upward movement of water vapor in chernozem soils in Russia. Chaptal (1934) recognized that soil moisture in the upper layers may be augmented by distillation from deep layers but did not regard the process as very important.

ADDITIONS AND LOSSES OF SOIL WATER

The moisture status in a soil body at any time represents a balance between additions and losses. The amount of water entering a soil is influenced by such factors as amount of precipitation, nature of the soil, character of the living and nonliving cover, and topography. Losses of soil water are influenced by gravity, evaporation, and transpiration.

ADDITIONS TO SOIL WATER

Precipitation is the principal source of water in soils. The amount which falls varies greatly from region to region, and within a given locality the amount varies during the course of

time. Annual precipitation is usually greater in forest regions than in grassland regions. The intensity of precipitation is also an important consideration in determining how much water enters the soil. Only when rainfall intensity exceeds the infiltration capacity of the soil can run-off occur. In addition to precipitation, some water reaches the soil in the form of water vapor from the atmosphere (Lebediev, 1924; Lebedeff, 1928; Chaptal, 1934).

The nature of the soil is highly important, since it influences the rate of infiltration of water reaching the ground surface. Schumacher (1864: 134-135) was perhaps the first to recognize that the permeability of soils to water is conditioned primarily by the noncapillary pore volume.² Soils having a large noncapillary pore volume are, as a rule, readily permeable to water (Burger, 1923; Joffe, 1932; Lutz, 1940; Free *et al.*, 1940).

Rate of infiltration is influenced not only by the total volume of noncapillary pores but also by their size and arrangement. Canals left by decayed roots, burrows of animals, cracks resulting from soil shrinkage, and the large pores between individual soil particles or aggregates all contribute to the noncapillary pore volume of soils. Wollny (1893) reported much higher percolation rates in soils having crumb structure than in soils that were structureless. An investigation carried out in coniferous plantations in New York revealed that the rate of infiltration through mull was 2.7 times greater than through the compact soil beneath mor (Heiberg, 1942). Infiltration is usually more rapid in coarse-textured than in fine-textured soils.

If snow falls on unfrozen ground, it may prevent the soil water from freezing, or at least the depth of frost penetration will be decreased, thus favoring continued permeability during the winter and early spring months. On the other hand, permeability may be greatly impaired if soil having low noncapillary pore volume becomes frozen before snow falls. In this connection it may be noted that Post and Dreibelbis (1943) found that the effect of soil freezing on retardation of percolation near Coshocton, Ohio, was greatest in heavy-textured and least in light-textured soils. Soil protected by forest cover does not freeze as early in the fall

² Noncapillary pore volume is the volume of the pores that are too large to permit appreciable capillary movement. These pores are emptied on the pF scale between pF 0 and the first flex point.

as soil that is bare, but it may remain frozen later in the spring. The important point is that, even when frozen, soil under forest cover is usually far more permeable than soil in open areas. This is explained by the fact that forest soils usually have many non-capillary pores that remain open after the water in the capillary pores has frozen. Augustine (1942) reported nearly normal infiltration rates on forest soil (Volusia gravelly silt loam) frozen to a depth as great as 4 in.

The rate at which water enters a soil is influenced by its initial moisture content. Fine-textured soils when dry may have shrinkage cracks and other noncapillary pores between the soil aggregates which permit rapid infiltration at the beginning of a storm. As the soil is wetted, the colloidal materials absorb water and swell, thus reducing the size of the pores. This accounts for the frequently observed fact that infiltration in fine-textured soils decreases materially with increasing moisture content. Infiltration in coarse-textured soils does not undergo such pronounced changes with differences in moisture content.

The incorporation of organic matter in mineral soils usually increases their permeability to water as a result of increased porosity. Investigations in the Boise National Forest in Idaho demonstrated clearly that the rate of absorption of water increased greatly with increased content of organic matter (Anon., 1940). Absorption, in terms of inches of water per hour, averaged 1.67 for soils having less than 2 per cent organic matter and 9.81 for soils having over 5 per cent. A decline in permeability of soils is commonly associated with, and in part caused by, a decrease in content of organic matter.

Vegetation cover affects the amount of water entering a soil in several ways. Substantial amounts of precipitation are intercepted by tree crowns and evaporated into the atmosphere. The amount thus prevented from reaching the soil varies with composition, age, and density of the stand and the season of the year. During a light rain of short duration interception is relatively larger than during very heavy or prolonged rains. Interception amounting to from 20 to over 30 per cent has frequently been reported for forest stands (Anon., 1940).

The organic debris which is contributed to the soil by vegetation plays a highly important role in determining the disposition of rainfall. It is frequently, but incorrectly, assumed that the

principal value of unincorporated organic matter lies in the fact that it can absorb several times its own weight of water. This high absorbent capacity is actually much less important than the influence exerted on infiltration and detention. Lowdermilk (1930) and others have emphasized that the beneficial effects of a cover of organic matter continue long after its absorptive capacity has been satisfied. Rain drops striking a bare-soil surface tend to disperse the soil aggregates. As the water moves downward, the soil particles held in suspension are filtered out, clogging the pore space. As a result the porosity of the surface soil layer is reduced, and infiltration is greatly decreased. If a cover of organic matter is present, conditions for infiltration are far more favorable. The impact of the raindrops is broken by the litter, and the water moves gently into the mineral soil without agitating the soil particles and causing them to come into suspension. Another important effect of unincorporated organic matter is a reduction in the rate of the overland flow of water. Friction increases with increasing subdivision of the flow by the leaves, needles, twigs, bark, and other organic debris on the mineral soil. After periods of rainfall in excess of the infiltration capacity some water remains on the ground as residual detention. A large proportion of this residual detention enters the soil, thus reducing total run-off and increasing total infiltration. Wollny (1888), Johnson (1940), and Anon. (1940), together with many others, have recognized the favorable effects of organic matter on infiltration.

Although a cover of organic matter on the mineral soil usually favors infiltration, it should be recognized that exceptions may occur. Ramann (1906), Engler (1919), and Wittich (1938) reported unfavorable effects when thick layers of organic matter, resistant to wetting, formed a continuous cover over the mineral soil. It is probable that a fibrous mor type of humus layer is less desirable from the standpoint of rapid infiltration than are other mor or mull types.

During heavy rains less water enters the soil on steep slopes than on relatively level areas. Run-off tends to increase with increase of slope. On the other hand, soil which is situated at the base of a slope or in a valley may receive water which has moved in laterally. Most forest soils are characterized by micro-relief features which owe their origin to the uprooting of trees.

Substantial depression storage for water is provided in the pits caused by upheaval of the root systems; these depressions must be filled to overflow levels before surface run-off can occur. The water in the depressions is held until it can enter the soil. Opportunities for depression storage are usually greater in forest soils than in agricultural soils.

The infiltration capacity of a forest soil is usually decreased as a result of cultivation, excessive grazing, or repeated burning. Unfavorable results are most pronounced in heavy-textured soils. Investigations have shown that the permeability of good forest soils generally exceeds that of cultivated fields or pastures (Engler, 1919; Burger, 1923; Auten, 1933; Stewart, 1933; Anon., 1940). In tests made on seven soil types in the Ozark region of Missouri it was found that annual burning reduced infiltration an average of 38 per cent in comparison with that in soils protected for approximately 5-6 years (Arend, 1942).

LOSSES OF SOIL WATER

A substantial part of the water which enters a forest soil is not held but is moved downward by gravity. Each soil has a limited capacity to hold water against the force of gravity; and, when this capacity is exceeded, the surplus appears as gravitational water. If precipitation is added to a sandy soil and a clay soil, both of which were initially dry, gravitational water will flow out of the sandy soil first. This fact is explained by the relatively low field capacity of the sand.

Losses of soil water by evaporation may be substantial. Of the many factors that influence evaporation the following appear most worthy of mention:

1. Moisture content of the atmosphere.
2. Wind velocity.
3. Soil cover, both living and nonliving.
4. Nature of the soil, including moisture content and temperature.

Moisture losses resulting from evaporation usually increase as the relative humidity of the atmosphere decreases. This fact does not mean, however, that any rigorous proportionality exists between relative humidity and evaporation rate. A more precise

statement would be that, neglecting the influence of wind, the rate of evaporation is proportional to the difference between the vapor pressure at saturation at the evaporating surface and the actual vapor pressure in the air above. Evaporation losses generally increase with wind movement, because the air immediately above the evaporating surface is continually being replaced by air having a lower vapor pressure. European investigators have attributed much importance to the harmful influence of wind in causing excessive moisture losses from the soil near the edge of forest stands. These losses may be minimized by maintaining good vertical closure at the forest borders.

Soil cover, both living and nonliving, reduces evaporation losses. This fact is to be expected in view of the effect of cover on wind velocity and soil temperature. There are many investigations which demonstrate that the evaporating power of the air in a forest is less than in the open (Anon., 1940). However, data relative to the influence of a forest on actual evaporation from the soil are scanty. Ebermayer (1873) reported that forest cover alone, with no organic debris on the soil, decreased evaporation by about 62 per cent in comparison with the loss from bare soil in the open. Bühler (1918: 341) placed the reduction at 53-75 per cent.

A cover of organic matter was reported by Eser (1884: 72-73) to reduce evaporation by 85-90 per cent; values by Bühler (1918) indicated a reduction of 30-35 per cent. Heyward (1939) found that upper soil layers in unburned longleaf pine forests contained more moisture than did the soil in burned forests. The greater amount of moisture in the unburned areas appeared to be due in part to the mulching effect of the rather thick rough which developed in the absence of fires. Other investigators who have reported a reduction in evaporation as a result of a cover of litter or slash include Albert (1924), Weisbrod (1935), and Stålfelt (1937). It was pointed out by Stålfelt that immediately after a rain, when the surfaces exposed to the atmosphere are still wet, more water is lost from areas covered with litter than from bare soil.

The nature of the soil itself influences water losses by evaporation. Fine-textured soils lose more than coarse-textured ones. Eser (1884: 72-73), Helbig and Roeszler (1921), and others have

found evaporation losses lower in soils having crumb structure than in single-grained soils. Evaporation from wet soils is extremely rapid; generally the losses are greater than from free water surfaces (Warington, 1900: 109; Stålfelt, 1937). As the moisture content of the surface soil decreases, there is a pronounced decrease in the rate of evaporation loss. This is readily understandable in view of the slowness of movement of capillary water. Veihmeyer and Hendrickson (1934) stated that in California very little water that passes below the upper foot of soil is subsequently lost by evaporation.

Evaporation usually increases with a rise in soil temperature. It is common experience that soils on south slopes are drier than those on north slopes. This observation is confirmed by the experimental data of Eser (1884) and others, who have shown that evaporation losses on south slopes are much greater than on north slopes.

The available evidence, which admittedly is scanty and unsatisfactory, indicates that losses of soil water as a result of transpiration by forest trees and other vegetation are very great. Preliminary results reported by Hoover (1945) indicate that in the southern Appalachians, near Franklin, North Carolina, a forest stand with a dense understory of shrubs annually transpires from 17 to 22 in. of water. Transpiration losses, which certainly are larger than evaporation losses, are influenced by factors such as climate, soil conditions, and nature of vegetation. Most of the available information on transpiration losses has been reviewed by Zon (1927), Raber (1937), and Anon. (1940). Mention may be made here of the differences in the water content of soils supporting forest and those which are bare. It has been observed repeatedly that during the vegetative period the forest-supporting soils are drier than the bare ones (Ebermayer, 1889; Craib, 1929; Vyssotzky, 1932; Ijjász, 1938). Dengler (1935: 282) reported that initially wet soils stocked with spruce dry out as the stands develop to the pole stage.

SOIL MOISTURE CONSTANTS

Many investigations have been devoted to the determination of "constants" or "equilibrium points" for use in characterizing

soil-moisture relations. Among the most widely used constants are the following:

1. Hygroscopic coefficient.
2. Moisture equivalent.
3. Wilting coefficient.
4. Field capacity.
5. Sticky point.

The approximate positions of some of these constants on the soil-moisture scale are shown diagrammatically in Figure 35, page 299.

The *hygroscopic coefficient* is supposed to mark the upper limit of the hygroscopic moisture range (approximately pF 4.5). The method usually employed in determining this value has been to expose for 24 hours a thin layer of dry soil to an atmosphere saturated with water vapor. It is now recognized that the "hygroscopic coefficient" so obtained is not an equilibrium point at all. A soil exposed to an atmosphere saturated with water vapor will continue to absorb moisture until it is saturated (Puri, 1925; Keen, 1931: 214). Puri (1925) and Puri, Crowther, and Keen (1925) advised that the hygroscopic coefficient be replaced by a measurement of the moisture content of a soil in equilibrium with an atmosphere of 50 per cent relative humidity. Olmstead and Smith (1938: 910) also recognized that exposure of a soil sample in a saturated atmosphere gives unsatisfactory results and suggested that it is better to expose the sample for a period of 5 days in an atmosphere only 99 per cent saturated with water vapor. Even the value obtained in this way is not an equilibrium point; the sample will continue to take up small amounts of moisture for weeks. Consequently, values obtained as suggested by Olmstead and Smith are probably slightly below the true hygroscopic coefficient. It may also be pointed out that the hysteresis effect is a factor which influences the moisture content of a soil in equilibrium with an atmosphere of a given vapor pressure. A soil which has reached an equilibrium as a result of drying contains more water than one which has reached an equilibrium as a result of wetting. Olmstead and Smith (1938) refer to the hygroscopic coefficient as "hypothetical and elusive"; Keen (1931: 213) was even more critical and stated that it had no real existence.

Hygroscopic moisture values have found wide use in spite of their unsatisfactory nature. In general, the highest values are

reported for soils having a high content of colloids. It should be borne in mind, however, that not only the amount of colloid but also its nature and the adsorbed cations influence hygroscopicity; with a decrease in the silica-sesquioxide ratio there is a decrease in hygroscopicity (Baver, 1940: 71). Consequently, methods for determining the surface area of soil particles, such as that of Mitscherlich (1923: 72-77) are open to serious question. Soils containing no moisture in excess of that indicated by the hygroscopic coefficient are incapable of supplying water to plants.

Briggs and McLane (1907, 1910) introduced the term *moisture equivalent* to designate the percentage of water retained by a soil when the moisture content was reduced by means of a constant centrifugal force until it was brought into a state of capillary equilibrium. They regarded the moisture lost by centrifuging at a speed equal to a force 3000 times gravity as that removed from the larger capillary pores. Later Briggs and Shantz (1912) used a centrifuging speed equal to 1000 times gravity in their determinations; this is the usual practice today. The technique of determining the moisture equivalent has been investigated and improved by Veihmeyer, Israelsen, and Conrad (1924) and Veihmeyer, Oserkowsky, and Tester (1928). In 1929 Bouyoucos introduced a new method for determining the moisture equivalent; by using vacuum pressure force, rather than centrifugal force, the need of a centrifuge was obviated. This method appears desirable only when a centrifuge is not available.

Moisture-equivalent values usually have been based on soil material that passed a 2-mm. sieve. The use of samples of undisturbed soil ("soil in place") appears to be better practice in forest soil work. Davis and Adams (1928) and Duncan (1939) found that the moisture equivalent of soil-in-place samples was usually lower than that of corresponding samples of sieved soil.

The principal initial use of moisture-equivalent values was to estimate the wilting coefficient. Its use for this purpose has declined in recent years for reasons which will be mentioned later. Meanwhile, a new use has developed, namely, as an approximation of the field capacity. The possibility of employing moisture-equivalent values for this purpose has been recognized by various soil scientists. For example, Shaw (1927) pointed out that the "normal moisture capacity" (practically synonymous with field capacity) of medium-textured soils is approximately the same as

the moisture equivalent. Veihmeyer and Hendrickson (1931), Olmstead and Smith (1938), and Browning (1941) also reported that the moisture equivalent is approximately equal to the field capacity in many soils; in clays the moisture equivalent is slightly higher, and in sands considerably lower, than the field capacity. More recently Edlefsen and Anderson (1943: 33) have expressed a similar view, stating that the moisture equivalent is a convenient, and for most soils a fairly accurate, method of evaluating field capacity. Olmstead and Smith (1938) indicate that the hygroscopic coefficient of many soils is about 0.37 times the moisture equivalent.

Several investigators have found a broad relation between the moisture equivalent and mechanical analysis data (Smith, 1917; Middleton, 1920; Veihmeyer and Hendrickson, 1931). Middleton expressed the relation between the moisture equivalent and the percentages of sand, silt, and clay as follows: moisture equivalent = $0.063 \text{ sand} + 0.291 \text{ silt} + 0.426 \text{ clay}$. Moisture-equivalent values increase with increase in content of colloidal material; they are also influenced by the chemical composition and replaceable bases of the colloids.

The pF value at the moisture equivalent is approximately 2.7 for nearly all soils. This definite energy relationship increases the value of the moisture equivalent in characterizing the moisture status of a given soil.

The *wilting coefficient* was defined by Briggs and Shantz (1912) as the moisture content of the soil, expressed as a percentage of the dry weight, at the time when the leaves of the plant growing therein first undergo a permanent reduction in their moisture content as a result of a deficiency in the moisture supply of the soil. Sunflowers are usually used as the test plants. A permanent reduction in the moisture content of the wilted basal leaves is indicated when they fail to regain turgor after the culture is placed in an atmosphere approximately saturated with water vapor. Attention is directed to the fact that the wilting coefficient, as originally defined by Briggs and Shantz, was based on the amount of soil moisture present when the test plants first underwent permanent wilting. The term was not originally restricted, as is sometimes assumed, to values obtained indirectly (moisture equivalent divided by 1.84). Misunderstanding on this point appears to have been responsible for the tendency among some writers to

restrict use of the term *wilting coefficient* to indirectly determined values and to employ other expressions, such as *permanent wilting percentage* or *wilting percentage*, to values obtained directly by means of test plants.

It may be pointed out that plants continue to absorb water from the soil even after the wilting coefficient has been reached. After the soil moisture has been reduced to the wilting coefficient, water uptake by the plant is, of course, restricted. Wilting, which starts with the basal leaves, progresses upward until finally even the apical leaves are affected. When this stage of moisture reduction is reached, the plant soon dies, unless supplied with water. Taylor *et al.* (1934) have employed the expression *ultimate wilting point* to designate the soil-moisture condition when all the leaves of the test plant are wilted. They also developed the concept of the *wilting range*, which is defined as the range in moisture content of the soil between the wilting coefficient and the ultimate wilting point. The researches of Taylor *et al.* (1934) and Furr and Reeve (1945) have shown that soil moisture within the wilting range may be of considerable ecological importance in enabling plants to survive prolonged periods of drought. Furr and Reeve reported that from 11 to 30 per cent of the available soil moisture (moisture between the ultimate wilting point and the moisture equivalent) is held within the wilting range. Although the wilting coefficient has been generally regarded as marking the boundary above which water is available to plants and below which water is unavailable, it now appears that the amount of soil moisture within the wilting range is sufficiently great to merit attention.

The view has been frequently expressed that the content of soil moisture at the time of wilting differs with different plants. This belief, however, does not seem to be warranted. Veihmeyer and Hendrickson (1934) found the permanent wilting percentage remarkably constant for different species growing in the same soil. Furthermore, they and others have found that climatic conditions do not materially affect this value.

The wilting coefficient is highly important in considerations of the amount of water readily available for plant use. A clay soil having a wilting coefficient of 15 and a water content of 10 per cent is physiologically dry, but sand with a wilting coefficient of 2 and a water content of 4 is physiologically moist. Veihmeyer

and Hendrickson (1934) concluded that plants can, with equal facility, use water from the field capacity to about the permanent wilting percentage. Plants wilt when the pF value of the soil is about 4.2. This fact means that, if an accurate curve of pF and moisture content is available for a soil, the moisture content at which plants will wilt can be read from the graph.

The fact that the addition of organic matter to mineral soils results in greatly increased water-holding capacity has led many writers to the conclusion that the supply of available water is also increased. The investigations of Feustel and Byers (1936) indicated that the moisture economy was benefited by additions of organic matter to quartz sand. Favorable effects on the supply of available moisture were less in loamy fine sand soil, and little or no advantage was apparent in a clay loam. It should be borne in mind that, although organic matter can absorb large quantities of water, its wilting coefficient is also very high. Wilting coefficient values of more than 80 have been reported for organic matter.

Appropriate methods of determining directly the wilting coefficient by means of test plants are described by Briggs and Shantz (1912), Taylor *et al.* (1934), Feustel and Byers (1936), and Furr and Reeve (1945).

An indirect method for arriving at the wilting coefficient was developed by Briggs and Shantz (1912). They proposed to divide the moisture equivalent by the factor 1.84, but during the years that followed many criticisms were directed against this proposal (Alway, 1913; Keen, 1922; Veihmeyer and Hendrickson, 1934; Duncan, 1939). Veihmeyer and Hendrickson found that the ratio varied from 1.39 to 3.82, and Duncan reported variations from 1.57 to 5.65. It is evident that the factor 1.84 should not be used for all soils.

Bouyoucos (1938b) proposed a new indirect method for determining the wilting coefficient of soils. This method is based on the fact that at or above the wilting coefficient the soil moisture film is sufficiently thick to cause soil particles and granules to cohere when pressed together, whereas below the wilting coefficient cohesion is lacking. This method has not come into general use.

Field capacity is a term used by Veihmeyer and Hendrickson (1931) to indicate the amount of water held in a soil after excess

gravitational water has drained away and after the rate of downward movement of water has materially decreased. The field capacity appears to be similar to the "maximum capillary water capacity" of Widtsoe and McLaughlin (1912) and the "normal moisture capacity" of Shaw (1927). In many soils the field capacity is about equal to the moisture equivalent. The pF value for soils at the field capacity is 2.7 or somewhat less.

The field capacity of soils in a given region is influenced by a number of factors, among which texture, organic matter content, structure, and depth of the soil body to ground water may be mentioned. Texture is by far the most important of these factors; both the field capacity and the amount of water readily available to plants are, for the most part, controlled by texture. The addition of organic matter increases the field capacity of coarse-textured soils, but the effects are negligible in clays. It should be recognized that both texture and content of organic matter may differ in the different horizons of a soil. For example, the A_2 horizon of a podzol usually contains less inorganic and organic colloidal material than the B_2 horizon. The field capacity of the A_2 horizon is much less than that of the B_2 horizon.

Field capacity, based on either volume or weight, may be determined directly for soils *in situ*. Soils in which drainage is unobstructed are generally considered to be at field capacity 1-5 days after a prolonged rainfall or irrigation. Many workers regard 2-3 days as adequate time for the soil moisture to attain the desired energy level. Actually the time required varies in different soils; coarse-textured soils drain to their field capacity more quickly than do fine-textured soils. For practical purposes it may be assumed that a soil has reached its field capacity when samples taken 8-12 hours apart, after a prolonged rain or irrigation, show essentially the same moisture content. Depth of wetting is an important consideration, as has been shown by Colman (1944), working with mountain soils in the San Dimas Experimental Forest in southern California. In the soils investigated, wetting to a depth of 12-30 in. was required before the moisture content of the surface layers attained field capacity.

Moisture content at field capacity, as has been indicated, may be expressed as a percentage of either dry weight or volume of soil. It is apparent that moisture content based on dry weight

will be greatly influenced by the specific gravity of the soil material. It is equally apparent that the important consideration, from the viewpoint of the plant ecologist, is the amount of moisture in a given volume of soil rather than in a given weight. Recognition of these facts has led to increasing use of volume as a basis for expressing soil moisture percentages.

The literature abounds with references to laboratory measurements of the capacity of soils to retain water against the force of gravity. Terminology varies greatly, as do also the methods employed (Kopecky, 1914, 1928; Burger, 1923; Köttgen, 1926; Davis and Adams, 1928; and Gulisashvili and Stratonovitch, 1935). Determinations usually have been based on the moisture content of relatively short columns of undisturbed soil that were first saturated and then allowed to drain. Although details of the methods vary, the results have one feature in common, namely, they indicate a moisture content substantially in excess of the field capacity. Information supplied by laboratory measurements of the moisture content of short soil columns after drainage appears to have only limited application. Data supplied by field-capacity determinations on soil *in situ*, pF curves, or the moisture equivalent appear far more useful.

The *sticky point* was advocated by Hardy (1923) as a useful measure of the water relations in soils. Investigation of this constant by Keen and Coutts (1928) confirmed the views of Hardy concerning its value in characterizing soil-moisture relations. The sticky point is believed to measure the water adsorbed by the colloidal material and the water held in the small interstices of the soil. The measurement is correlated with other soil characteristics, such as loss on ignition and content of colloidal material. In the majority of mineral soils the sticky point increases as the texture becomes finer. However, sandy soils with a high content of humus also give high values. Interrelationships of the sticky point and other soil properties have been investigated by Bodman and Perry (1931). For methods the authors mentioned above and the discussion of Keen (1931) and Olmstead (1937) should be consulted.

FORMATION OF ICE LAYERS AND FROST HEAVING

It is a well-known fact that the heaving which frequently occurs in soils is much greater than that attributable to the

change in volume resulting from freezing of the water initially present. Detrimental frost heaving, that is, heave of several inches, is related to the formation of ice layers in the soil, which in turn depends on the amount of water available for freezing at a given location.

FORMATION OF ICE LAYERS

As freezing progresses downward from the surface, water may be drawn from the underlying, unfrozen soil to the base of the frozen layer. Segregation of ice layers, with attendant heaving, results (Bouyoucos, 1923; Beskow, 1935; Nömmik, 1940; Taber, (1943). A significant desiccation of nonfrozen soil layers adjacent to frozen layers has been shown by Nömmik.

Texture, structure, and amount of water available all influence the formation of ice layers in soils. Segregation of ice appears to be most frequent in silts and loams, less common in heavy clays, and rare in coarse sands and gravels (Beskow, 1935; Hogentogler, 1937: 147-151). Coarse crumb structure and the presence of large noncapillary pores in a soil impede the upward movement of water, thus inhibiting the formation of ice layers (Beskow, 1935). The formation of ice layers requires that water be conveyed to the zone of crystallization. Consequently, there must be a large amount of mobile water in the soil itself or else a capillary connection with the ground water. Conditions for ice formation and frost heaving are most apt to be present in relatively fine-textured soils having low noncapillary porosity and a high water-table. Spaeth and Diebold (1938) reported serious heaving, caused by the formation of thick layers of ice, in soils of open fields where the water-table was close to the surface.

FROST HEAVING

The increase in volume of water on freezing is principally in the direction of heat loss. In soils the direction of heat loss and the resulting increase in volume are mostly vertical.

In some situations considerable damage to seedlings is caused by heaving. Roots are broken, and in severe cases the young trees are literally heaved out of the ground (Haasis, 1923). Damage is most common in open areas where the soil is bare or

has only a scanty cover. This condition may exist in old fields or in areas that have been eroded. A cover of organic debris or snow greatly lessens the likelihood of destructive frost heaving. Heaving is also of importance in causing soil creep on slopes which are bare or which support only a sparse cover of vegetation.

ECOLOGICAL SIGNIFICANCE OF SOIL MOISTURE

As a result of transpiration, large amounts of water in the form of water vapor pass into the atmosphere from the leaves of trees and other plants. This water loss must be balanced by uptake of water from the soil. For normal growth the ratio of water uptake to water transpired must be greater than unity. It appears that the water requirement of plants, that is, the units of water required for each unit of dry matter produced, may vary from about 300 to well over 500 (Raber, 1937). The water requirement of plants growing in soils of low fertility is higher than that of plants growing in fertile soils. Continuity of water supply for forest trees is fully as important as the total amount.

In many regions the occurrence of forest types is largely controlled by the supply of water. For example, the lower altitudinal limits of the various forest types in the central Rocky Mountain region and in the Southwest are for the most part determined by the water supply. In semiarid sections the amount of available soil moisture frequently determines whether forest or grassland vegetation dominates.

Usually site quality improves with increasing amounts of available soil moisture. There is, of course, a limit to the amount of soil water which is desirable; and, if this limit is exceeded, unfavorable conditions for plant growth result because of deficient aeration and related phenomena. An excessive amount of water may be quite as unfavorable for tree growth as a deficiency. Wiedemann (1938) pointed out that growth of beech in the usually dry climate of Hessen dropped about 20-30 per cent during a period of low rainfall from 1890-1896. During this same time, however, growth of beech in the usually moist climate of Schleswig-Holstein increased; in other words, growth in Schleswig-Holstein was greater during periods of low rainfall than when rainfall was high. In humid regions coarse sandy soils with deep water-tables usually represent poorer sites than

do soils of medium or fine texture. On the other hand, tree development in dry regions is commonly best on sandy, gravelly, or rocky soils (Pearson, 1931; Scholz, 1935). Pearson reported that in the Southwest the effects of drought are usually first apparent on clay soils.

In an evaluation of the moisture relation of soils consideration should always be given to the nature of the deeper soil horizons and underlying strata. Layers of fine-textured material lying several feet below the soil surface may be highly important, as von Kruedener (1934) and Pearson and Marsh (1935) have pointed out. Tamm (1920: 298) stated that soil moisture was commonly deficient in the pine-heath areas of northern Sweden. Podzolization, with the formation of *Ortstein*, was held to be beneficial in these areas because water relations were improved.

The presence of a water-table near the surface is known to have a profound influence on the productivity of soils. Albert (1925) pointed out that a high water-table could compensate for the low field capacity of coarse sandy soils, with the result that they were very productive. Hartmann (1930) stated that a water-table as much as 16 ft. below the surface was observed to influence the growth of pine. Stoeckeler and Sump (1940) found that direct seeding of conifers was feasible on sand-plain areas in the Lake States where a water-table that did not fluctuate greatly was 2-5 ft. below the surface. In such situations they invariably found more available moisture in the surface 6 in. of soil than was found in areas where the water table was deeper than 6 ft. In addition to the depth at which ground water occurs, the content of oxygen and nutrient elements is also important. The oxygen content of the ground water generally increases with increasing rate of movement. A very high water-table, particularly if the ground water is not undergoing any appreciable horizontal movement, may result in a physiologically shallow soil (Dengler, 1935).

The importance of root competition for soil moisture in forest stands has been demonstrated by a number of investigators (Fricke, 1904; Toumey and Kienholz, 1931; Korstian and Coile, 1938). When root competition was eliminated by trenching, an increase in soil moisture content was usually observed. In response to the increased soil moisture, tree seedlings and herbaceous plants appeared. Kmonitzek (1930) reported that

underplanting spruce and beech in a pine forest in Germany resulted in decreased soil moisture as a result of root competition. This condition was reflected in poorer growth of the overwood.

Competition for soil moisture may become exceedingly keen as the trees in dense stands approach the pole-wood stage (Albert, 1915; Dengler, 1935: 282). The application of thinnings or other cuttings represents the only means at the forester's disposal for alleviating this condition. Thinnings affect the soil-moisture relations by decreasing the transpiration drain on the soil and also by enabling a larger percentage of the precipitation to reach the soil. The silviculturist must realize that sites which tend toward dryness cannot support as dense stands as those which have a good water supply. Failure to recognize this fact and maintenance of too large a growing stock on dry situations are almost certain to result in substantial losses during drought years.

METHODS FOR MEASURING INFILTRATION

Many techniques have been developed for the measurement of infiltration of water into soils. Some are simple and employ inexpensive equipment, but others require elaborate and expensive facilities. Two categories of techniques may be recognized: (1) infiltration is regarded as equivalent to the water applied when run-off is prevented, and (2) infiltration is regarded as equivalent to the difference between water applied and run-off when the latter is permitted.

The earliest, and possibly most widely used methods in the first category, involved measurement of the rate at which water entered the soil when run-off was prevented by a cylinder or box. Kopecky (1914, 1928) regarded the "relative permeability" of a soil as the volume of water which in 24 hours would pass through a soil column having a cross-sectional area of 10 sq. cm. and a height of 10 cm. The soil was brought to its field capacity before the test was made. A somewhat different technique was employed by Burger (1923). A cylinder having a cross-sectional area of 100 sq. cm. and a height of 20 cm. was driven into the soil to a depth of 10 cm. One liter of water was then added, and the time required for infiltration was recorded. The technique proposed by Burger has been used in America by Stewart (1933), Lutz (1940), and others. Additional methods involving the use

of cylinders or boxes to prevent run-off are reviewed by Davidson (1940).

Kohnke (1939) described an apparatus for measuring infiltration which consisted of 16 rectangular compartments. This apparatus was driven 1 or 2 in. into the ground to prevent run-off, and water was added to each of the compartments. Infiltration rates were based on the average of the data obtained from the 4 internal compartments. The water added to the outer or buffer compartments was believed to restrict the lateral subterranean flow from the internal compartments.

In recent years many techniques of the second category have been developed. Pearse and Woolley (1936) determined the absorption and run-off of water, applied at a known rate, on plots 1 ft. square. The North Fork infiltrometer, described in detail by Rowe (1940), makes it possible to sample areas 1 ft. wide and 2.5 ft. long. Relatively even distribution of simulated rainfall at intensities of from 0.75 in. to more than 10 in. per hour is obtained by means of a sprinkler system. Workers in the federal services have developed apparatus to simulate natural rainfall of varying intensities on plots as large as 0.01 acre. The work of Horton (1937), Diebold (1942), and Wilm (1944) and the compilation of Davidson (1940) should be consulted for a more complete review of the various methods which have been used.

LYSIMETERS

Movement of water through the soil may be investigated by means of lysimeters. These devices permit measurement of the percolation of water through soils and determination of the constituents removed in drainage. Lysimeter installations may utilize either natural soils *in situ* or soil material which has been prepared and placed in the lysimeter in some particular manner. The use of disturbed soil in lysimeters has been strongly criticized by Joffe (1929, 1932).

Ebermayer (1879) was perhaps the first to use lysimeters to measure water movement through undisturbed soil in the forest. A pit was opened, and in one of the walls, at a depth of about 4.5 ft., a tunnel was prepared. A metal lysimeter was placed in the tunnel to permit measurement of the water which percolated through the overlying soil body. Joffe (1929, 1932) has used

a similar lysimeter with satisfactory results. Use of this technique permits measurement of the water percolating through the various soil horizons. Lysimeters of the shallow tank and pan types have been employed by Lunt (1937, 1941) in his investigations of the movement of water and nutrient constituents through the unincorporated organic matter, bare surface soil, and normal surface soil (including unincorporated organic matter) under forest stands.

In recent years the Soil Conservation Service has installed near Coshocton, Ohio, automatic recording lysimeters utilizing soil in place. The horizontal projection dimensions of each unit are 6.22 by 14 ft.; each lysimeter is 8 ft. deep. An automatic weighing device was developed which is capable of recording at stated intervals the total weight, within 5 lb., of the 130,000-lb. lysimeter and soil mass (Garstka, 1938). The Forest Service has established a large number of lysimeters in the San Dimas Experimental Forest in Southern California. They vary in size from metal tanks having a capacity of about 300 lb. of soil to large concrete lysimeters having a surface area of 5 milacres (10.5×21 ft.) and a depth of 6 ft. These installations are designed to supply information on the water cycle.

A comprehensive survey and discussion of lysimeters are given by Kohnke, Dreibelbis, and Davidson (1940).

MEASUREMENT OF SOIL MOISTURE

Methods for measurement of soil moisture may conveniently be considered in two groups: (1) those involving removal of a sample from the soil body, and (2) those in which the moisture of soil is measured *in situ*.

METHODS INVOLVING REMOVAL OF A SAMPLE FROM THE SOIL BODY

The most widely used method of determining soil moisture content is to dry a sample to constant weight in an oven having a temperature of 105° – 110°C . Soil moisture is commonly expressed as a percentage of dry weight, but in ecological work values based on volume are preferable. Securing such values, of course, necessitates use of soil samples having a known volume.

Several other methods that are more rapid than oven drying

have been proposed. Bouyoucos (1931) developed a technique that involved measurement of the change in density of alcohol when placed in contact with the moist soil particles. Later Bouyoucos (1938a) described a field method in which the soil moisture was evaporated by burning alcohol. Neither of these methods has come into general use.

All methods of soil-moisture determination that involve removal of samples from the soil body have the disadvantage of being time-consuming. The sample must be collected and weighed, then dried and reweighed. When changes in soil moisture are being studied over a period of time, another difficulty confronts the investigator. Successive samples cannot be taken from identical positions; new positions must be utilized each time samples are collected. This situation introduces an important source of error if the soil being investigated lacks uniformity.

METHODS FOR MEASUREMENT OF THE MOISTURE CONTENT OF SOIL *IN SITU*

Plant scientists, particularly those concerned with natural vegetation types, have long recognized the need of a technique whereby moisture changes in undisturbed soil can be followed throughout the growing season and during successive years. Although the problem has been receiving attention for nearly a half century, it is only recently that promising methods have been developed. Some of the recently proposed methods are still in the experimental stage, but others are perfected to a point where they give reasonably satisfactory results under field conditions.

Shaw and Baver (1939, 1940) used heat conductivity as a measure of soil moisture content. The method gives excellent results under laboratory conditions, but in field tests it does not permit satisfactory measurement of the moisture in the soil (Cummings and Chandler, 1941). In the field it is almost impossible to secure and maintain adequate contact between the soil and the heating unit as originally designed. Another disadvantage of this procedure is that a calibration curve must be prepared for every soil investigated, since each soil has its own characteristics of heat conductivity. Johnston (1942) modified the heater units of the Shaw and Baver apparatus by adding

jackets of plaster of paris. When placed in two very different soils, these jacketed heater elements showed comparable readings at field capacity; at the permanent wilting percentage comparable readings were again obtained, but these differed, of course, from the readings at field capacity. If the favorable results obtained in the initial tests, conducted in pots of soil with sunflowers, are confirmed by field experiments, the method will merit careful consideration by forest research workers.

It has been known for many years that the electrical resistance of soil varies with moisture content, and methods of moisture determination based on this fact have been proposed. In the early methods conductivity was measured between two electrodes placed directly in the soil. Results were usually unsatisfactory, however, because of erratic contact resistance between the electrodes and the soil and because of variations in the salt content of the soil solution.

These difficulties have been minimized in the new technique developed by Bouyoucos and Mick (1940). Two electrodes are still used, but instead of being placed directly in the soil they are embedded in a block of plaster of paris. With density, texture, compaction, and salt content fairly constant in the plaster of paris surrounding the electrodes, variations in electrical resistance reflect variations in the moisture content of the porous block. The moisture content of the block evidently varies with that of the soil, but only under conditions to be mentioned presently. In making moisture measurements the blocks are buried in the soil at the desired depths, and lead wires from the electrodes are brought to the surface. Readings are obtained by means of a modified Wheatstone bridge.

The researches of Anderson and Edlefsen (1942*a*, *b*), Edlefsen *et al.* (1942), and Edlefsen and Smith (1944) on the determination of moisture in undisturbed soil are of particular interest. These investigators found that electrodes in plaster of paris blocks, when placed in soils supporting actively transpiring vegetation, gave fairly dependable indications of soil moisture over the entire range of availability to plants. In soil lacking plants the blocks exhibited a marked lag in response to change in soil moisture, but this delay was not appreciable when growing plants were present. Lack of lag in the presence of plants may be due to maintenance by the absorbing roots of a steep moisture

gradient between the plaster of paris block and the immediately adjacent soil.

Measurements obtained by the Bouyoucos and Mick (1940) method have the distinct advantage of being correlated with the free energy or pF of the soil water. This fact means that results in different soils will be approximately the same for any given energy level. Thus, essentially similar measurements should be obtained in all soils having a moisture content equivalent to the permanent wilting percentage, or pF 4.2. Perhaps the greatest disadvantage of the Bouyoucos and Mick method is the fact that considerable time may be required for the establishment of equilibrium moisture conditions in the plaster of paris block and the adjacent soil. This lag is most serious in (1) very coarse-textured soils, (2) soils having relatively low moisture contents, and (3) soils devoid of growing plants. However, response of the blocks to changes in soil moisture, as has been pointed out, is generally satisfactory when plant roots permeate the soil body being investigated. Everything considered, the method proposed by Bouyoucos and Mick seems to be the most promising now available for measurement of the moisture content of soil *in situ*.

The dielectric method of Fletcher (1940) affords measurements that are correlated with the moisture content of the soil rather than with free energy or pF . The method has not been tested sufficiently to permit judgment of its practical value.

Tensiometers have been successfully used to measure the moisture content of soils *in situ* (Richards and Gardner, 1936; Wallihan, 1939; Lecompte, 1940; Richards, 1942; Veihmeyer *et al.*, 1943; Scofield, 1945). Among other uses they have been employed in forest tree nurseries to indicate when water should be supplied. Unfortunately tensiometers will not operate over the entire range of soil moisture available to plants. The upper limit of tension for these devices is about pF 3, which means that they are incapable of supplying readings at moisture contents approaching the permanent wilting percentage (pF 4.2). However, tensiometers have the advantage of measuring pF directly and are satisfactory in regions where rains occur rather frequently.

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Chapter 10

GENERAL CHEMICAL PROPERTIES OF FOREST SOILS

From a chemical viewpoint soils are composed of a large number of constituents, ranging from simple salts to highly complex inorganic and organic compounds. Many of the compounds occurring in soils are only poorly understood. Chemical processes continue unceasingly, with the result that soil composition is constantly changing. The soil may be likened to a vast chemical laboratory in which processes of great complexity and importance are going on continually.

Soil chemistry has an important bearing on the genesis of soils and on their physical and biological properties. It is a fundamental fact that plants require a source of certain mineral elements if they are to develop normally. Much of the interest and work in soil chemistry relates to the capacity of soils to supply adequate amounts of these essential mineral elements to plants.

Differences of opinion concerning the ecological importance of the chemical properties of soils are encountered. Some investigators, as has been pointed out, have taken the position that, as a rule, the physical properties of soils are of first importance and that chemical properties are of less significance. In specific cases this view may be entirely justified, but it is unwise to accept it as a valid generalization. Situations are certain to exist in which unfavorable physical conditions limit growth, and similarly other situations occur in which unfavorable chemical conditions become the most significant factor. In the opinion of Lang (1934) chemical properties of soils are most important near the northern limits of the thermal range of a species and least important near the southern limits. For satisfactory growth in forest stands excellence in both the physical and the chemical properties of the soil is essential. It appears that one reason for the casual consideration afforded chemical properties of forest soils by some writers is that trees use smaller amounts

of soil nutrients than do most agricultural plants. Another reason seems to be the fact that occasionally a soil poor in chemical nutrients may produce equally as good a timber stand as a soil which is rich in nutrients. This seemingly anomalous situation is generally explainable on the basis of compensating factors. For example, Wiedemann (1934: 17) has pointed out that in eastern Germany equally good pine stands may occur on gravels rich in lime but poor in fine sand and on fine sandy soils poor in lime. All soils studied which had a considerable content of lime at a depth of less than 6 ft. were good sites for pine; however, nearly all soils with a very low content of lime in the upper layers but with clay or ground water in the lower layers (at a depth of less than 6 ft.) were equally good. Kmonitzek (1930) reported that on sandy soils underplanting pine with beech and spruce seemed to affect favorably most chemical properties of the soil. Nevertheless, the underplanting influenced the pine overwood unfavorably, presumably because of competition for soil moisture. It cannot be emphasized too strongly that site quality, in so far as it is influenced by soil, depends on the integration of all soil properties, physical, chemical, and biological. The chemical composition of soils is influenced by the parent material, biological activity, climate, topography, and time.

NUTRIENT ELEMENTS

In addition to nitrogen, phosphorus, potassium, calcium, magnesium, sulfur, and iron it is now recognized that minute amounts of other elements, such as boron, manganese, zinc, and copper, are essential for many plants. Future research may add to this list. Plants are known to absorb many other elements from the soil, and it is possible that further investigations will demonstrate that at least some of them play a significant physiological role. Knight (1938) reported the identification of more than fifty chemical elements in the ash of leaves from a hickory tree growing in soil derived from pegmatite.

THE SOIL SOLUTION

The liquid that bathes the soil particles is usually termed the *soil solution*; some writers designate it as the *intermicellar liquid*.

It contains at least traces of all the elements present in the soil. The soil solution undergoes continuous fluctuations in amount, composition, and concentration; it is highly dynamic. Ions move from the colloidal particles into the soil solution and from the soil solution to the colloidal particles. Among the agents that operate to alter the amount, composition, and concentration of the soil solution the following may be mentioned: evaporation, rainfall, activity of microorganisms, absorption by plants of water and chemical constituents, and drainage.

Biologically the soil solution is highly important, since it, together with the colloidal particles, occupies a central position between the soil and the plant. Until recent years the view prevailed that ions had to be released to the soil solution before they could be taken up by plant roots. Evidence now points to a strong probability that plant roots in intimate contact with colloidal particles are able to take up adsorbed nutrient ions directly (Jenny and Overstreet, 1939; Overstreet and Jenny, 1940; Hoagland and Arnon, 1941; Overstreet *et al.*, 1942). In other words, there appears to be an exchange between cations held by the soil colloids and cations on the root surfaces. Plants evidently obtain nutrient ions partly from the soil solution and partly by contact mechanisms.

The ionic forms in which the more important nutrient elements are supplied to plants are as follows:

Nitrogen	NH_4^+ , NO_2^- , NO_3^-
Phosphorus	PO_4^{---} , HPO_4^{--} , H_2PO_4^-
Potassium	K^+
Calcium	Ca^{++}
Magnesium	Mg^{++}
Sulfur	SO_3^{--} , SO_4^{--}
Iron	Fe^{++} , Fe^{+++}
Manganese	Mn^{++} , Mn^{++++}
Copper	Cu^+ , Cu^{++}
Zinc	Zn^{++}
Boron	BO_3^{---}

For normal development of plants the soil not only must contain all the essential nutrient elements in available form, but also each element must be present in such concentration that a balance exists. This fact is well known to plant physiologists

and has recently been emphasized in connection with forest soils by Leiningen-Westerburg (1930) and Süchting, Jessen, and Maurmann (1937). Certain elements, if present in too high a concentration, influence growth unfavorably. In some soils the harmful influence appears to result from toxicity of elements present in excess; in others the elements present in excess appear to inhibit normal absorption of other essential ions.

GENERAL NUTRIENT REQUIREMENTS OF TREES

Relatively little specific information regarding the nutrient needs of forest trees is available. Of the many reasons which might be advanced to account for this lack of data complexity of the subject appears most important. All investigations of the nutrient requirements of plants present certain problems, but those arising in investigations involving forest trees are especially formidable. The aerial and subaerial organs of mature trees attain great size, generally making it impracticable to work with controlled conditions. Furthermore, many years elapse between the time of germination of a tree seed and maturity of the resulting organism.

Most of the data on the nutrient requirements of forest trees are based on ash analyses of tree parts. It should be recognized that data of this kind indicate the relative amounts of the various constituents actually absorbed by species in certain environments, but that they do not necessarily express requirements. If a large amount of a certain element is present in the soil in available form, it commonly will be absorbed in unusually large amounts (Süchting, Jessen, and Maurmann, 1937; and many others). Such unusually large absorption of certain elements is sometimes referred to as luxury consumption.

Various European investigators have presented data concerning the average annual consumption of mineral nutrients by forest stands (Ebermayer, 1876, 1882; Ramann, 1890: 52; Vater, 1920; Hackmann, 1930; Oelkers, 1930). Examination of these data indicates that the absolute and relative amounts of chemical constituents absorbed by different forest trees vary greatly. As a rule, hardwoods absorb larger amounts of most mineral nutrients than do conifers. The data concerning the ash composition of tree leaves which were presented in Chapter 6 substantiate this view. The fact that hardwoods usually require

a more fertile soil for satisfactory development than do conifers is understandable when the mineral nutrient consumption of species in the two groups is considered. The pines appear to make smaller demands on the mineral nutrient capital of soils than do any other trees. There seems to be a tendency for all species to absorb relatively large amounts of calcium. The uptake of this element usually exceeds that of magnesium, potassium, and phosphorus.

Nutrient consumption by forest trees varies with age. Ramann (1890: 105) stated that the maximum mineral nutrient requirement of forest stands comes in early or middle life and that requirements decrease with increasing age. On poor soils maximal requirements occur at a later age than on rich soils. Ramann regarded requirements as maximal in pine stands about 15-30 years old, in spruce stands about 30-50 years old, and in beech stands 40-60 years old. Morosov (1928: 104) also believed that the nutrient needs of forest trees increase up to an age of 20-35 years and then gradually decrease. Failures in plantation stands are known to occur at about this age and sometimes may be related to the period of maximum nutrient needs. Morosov (1928: 105) stated that, if trees are placed on soils not suited to them, they commonly do not show ill effects until they reach the pole-wood stage; the content of nutrients may be sufficient for normal development during early youth. Hackmann (1930) reported that the annual consumption of phosphorus by spruce (site class II) is greatest when the stands are about 30-50 years of age. During this period the annual consumption amounted to about 10.7-11.6 lb. per acre, thereafter it dropped to about 8.0-8.9 lb. annually.

In view of the fact that nutrient requirements of species vary it is obvious that no one "best" condition, applicable to all species and all conditions, can be specified. Likewise, it should be remembered that utilization of nutrients by trees depends not alone on their presence in the soil but also on other growth factors. The ability of a soil to supply nutrients and the capacity of vegetation to use them are quite different considerations.

NITROGEN RELATIONS

Various aspects of the nitrogen relations in forest soils have been considered in Chapters 5 and 6. In this section attention

will be directed to the source of soil nitrogen, amounts present, causes of losses, availability to plants, and requirements of forest trees.

The primary source of soil nitrogen is the atmosphere, of which about 80 per cent by volume consists of nitrogen in a free state. Atmospheric nitrogen, which is very inert, cannot be used directly by higher plants but must first be "fixed" or combined with other elements. In nature this process is accomplished by specialized groups of symbiotic and free-living microorganisms in the soil and by lightning discharges in the atmosphere. The work of the nitrogen-fixing microorganisms has been discussed in Chapter 5.

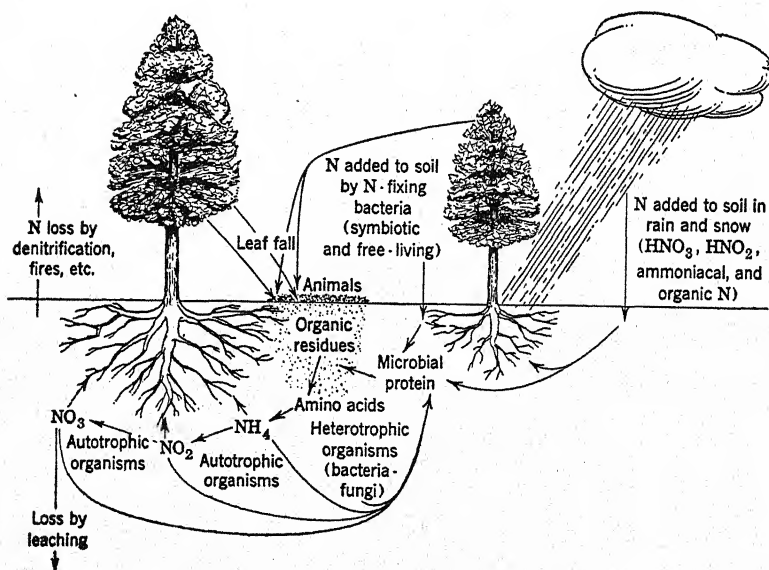


FIG. 36. Diagrammatic representation of the nitrogen cycle.

It is not possible to state definitely how much nitrogen is fixed by microorganisms in forest soils, but the amount must be substantial, particularly in stands containing legumes (Chapman, 1935). Probably it is five to ten times as great as that reaching the ground in rain and snow. F. W. Clarke (1924) and others have presented data which, although varying greatly, indicate that in temperate regions about 5-7 lb. of nitrogen per acre is annually added to the soil in precipitation. The nitrogen com-

pounds that are washed out of the atmosphere consist of nitrous or nitric acid, together with ammoniacal and organic nitrogen. Figure 36 is a diagrammatic summarization of the processes which are commonly referred to as the *nitrogen cycle*.

The amount of nitrogen present in soils differs from region to region. This variation is illustrated by the data presented in Table 36. Although the figures in this table must be regarded

TABLE 36. AVERAGE NITROGEN CONTENT IN VARIOUS SOIL REGIONS OF THE UNITED STATES

(After Schreiner and Brown, 1938)

Soil Region	Approximate Nitrogen Content in Surface 6 in., percentage	Average Nitrogen Content to a Depth of 40 in., percentage	Average Amount of Nitrogen per Acre to a Depth of 40 in., pounds
Gray-brown podzolic	0.05-0.20	0.05	6,700
Red and yellow Prairie	0.05-0.15	0.03	4,000
	0.10-0.25	0.12	16,000
Chernozem and chernozem- like	0.15-0.30	0.12	16,000
Chestnut	0.10-0.20	0.08	10,700
Brown	0.10-0.15	0.06	8,000

as approximate only, they do show that much higher values occur in regions occupied by tall-grass and mixed-grass prairie vegetation than in those characterized by forest.

Climate exerts an important influence on the amount of nitrogen in soils. Jenny (1929) showed that a correlation exists between the mean annual temperature and the average total nitrogen content. Nitrogen content decreases from north to south as mean annual temperature increases; for every 18°F. decline in temperature the average nitrogen content increases two or three times. The lowest values shown in Table 36 refer to the red and yellow soils of the South, which have developed under conditions of relatively high temperature. In a given climatic region the amount of soil nitrogen is influenced by such factors as kind of vegetation, nature of the soil, season of year, and amount of organic matter present.

Both nitrogen and organic matter usually decrease with in-

creasing depth below the ground surface. In podzol soil profiles the nitrogen content of the unincorporated organic matter is usually relatively high (about 1-4 per cent) and is followed by a sharp drop in the A_2 , an increase in the B , and a second drop in the C horizon. The decrease of nitrogen with increasing depth in other zonal soils is more regular.

Losses of nitrogen occur as a result of utilization of timber, removal or burning of organic debris, activity of denitrifying bacteria, and soil erosion, and in surface run-off and seepage water. The losses caused by removal of forest crops are poorly understood, but it is quite certain that they are much less than those resulting when agricultural crops are harvested. On an annual basis the removal of nitrogen by forest crops amounts to only about 5-25 per cent of the removal by agricultural crops. Although the immediate effect of a fire in a forest is a loss of nitrogen, it does not follow that the amount available to plants is decreased. It has been pointed out in Chapter 6 that burning frequently results in increased nitrification. Furthermore, the investigations of Heyward and Barnette (1934) and Wahlenberg *et al.* (1939) demonstrate that even *total* nitrogen is higher in longleaf pine soils that have been subjected to frequent fires than in those not thus influenced. Increased numbers of legumes on the burned areas, with greater fixation of atmospheric nitrogen by microorganisms, seem to be the explanation. It should not, however, be concluded that fires are always beneficial from the standpoint of nitrogen relations.

Denitrification may result in losses of nitrogen in soils that are poorly aerated. Nitrates are reduced to gaseous nitrogen by certain organisms in order to obtain oxygen. Soil erosion, a cause of serious losses of organic matter and nitrogen in agricultural soils, is of minor importance in the forest. Nitrates are readily soluble in water and, if not utilized by microorganisms or higher plants, are lost in the drainage water; under forest growth the amount of nitrate nitrogen lost is small. It is probable that the nitrogen brought down in rain and snow equals or exceeds that lost by leaching from soils supporting forest stands (Lunt, 1937).

From an ecological point of view the availability of soil nitrogen is highly important. Just as higher plants are unable to utilize directly atmospheric nitrogen, so also they are unable to utilize

nitrogen which exists in the form of complex compounds in the organic matter of the soil. Comparatively little is known concerning the nitrogen nutrition of forest trees, but from what has been learned about other plants it may safely be assumed that they depend principally on nitrate and ammoniacal forms. In addition they probably utilize some nitrite nitrogen and amino acids. On the basis of evidence now available there seems to be little ground for maintaining that nitrogen in nitrate form is superior to that in ammoniacal form. Hesselman (1927) regarded nitrates as highly beneficial, but in no sense essential, to the establishment of regeneration in Sweden. In northern forests and those at relatively high elevations in the mountains most of the available nitrogen appears to be in the form of ammonia with only small amounts of nitrate present (Aaltonen, 1926). Production of nitrates appears to be greater in mull types of humus layer than in mor types, but is not strictly confined to mull types.

The transformation of soil nitrogen occurring in complex, unavailable forms to forms which can be utilized by plants is a biological process influenced by many factors. The more important of these, most of which have already been considered in Chapters 5 and 6, are the following:

1. Very little nitrogen is liberated as nitrate until the carbon-nitrogen ratio of the organic matter has narrowed as a result of decomposition. The carbon-nitrogen ratio of the fresh organic debris is influenced by stand composition and soil fertility. Coile (1939) showed that the carbon-nitrogen ratio was lower, and the amount of nitrate formed greater, in an American sweet gum (*Liquidambar styraciflua*)-tulip tree stand than in a shortleaf pine stand. Kvapil and Némec (1925) reported that the ammonifying and nitrifying capacity of various kinds of humus was as follows: Spruce < beech < mixed spruce and beech.
2. Both ammonification and nitrification are inhibited by high soil acidity. Measures which reduce acidity favor nitrogen mobilization.
3. Deficient aeration, extremes of temperature, and low amounts of mineral nutrients and water in the soil inhibit the mobilization of nitrogen.
4. Nitrogen transformation in dense forest stands is frequently favored by thinnings. There is some evidence that the amount of nitrate decreases with increasing stand age.

5. In cool, humid regions exposure of forest soil by clearcutting is known to stimulate nitrification. Fire may also be used to advantage when dealing with mor humus types which do not nitrify readily as a result of exposure alone. Working of the soil to mix the mineral material and organic debris is another procedure known to favor nitrification. The use of fire and working of the soil are probably seldom necessary in American forests because the humus layer types usually react readily to influences such as thinnings and clearcutting.

The amount of available nitrogen in a given forest soil is not a constant value but fluctuates considerably. Under favorable conditions the combined amounts of nitrate and ammoniacal nitrogen in soils probably approximate 0.75–1.0 per cent of the total nitrogen. Utilization of nitrogen by vegetation and micro-organisms, time of leaf fall, and climatic conditions all have a bearing on the amount available at any given time. Fehér (1933) and Fehér and Frank (1936) have indicated that in forests of north-temperate regions a minimum nitrate content is commonly observed during the summer because of absorption by the trees.

Many investigators have regarded certain shrubby and herbaceous plants as useful indicators of nitrification in forest soils (Hesselman, 1917*b*; Bornebusch, 1931; Dengler, 1935: 150). Some of the species commonly believed to be "nitrate plants" are the following:

- Bristlestem hempnettle (*Galeopsis tetrahit*)
- Fireweed (*Epilobium angustifolium*)
- Groundsel (*Senecio* spp.)
- Herb Robert geranium (*Geranium robertianum*)
- Nettle (*Urtica* spp.)
- Red raspberry (*Rubus idaeus*)

The ammonifying and nitrifying capacity of forest soil material may be investigated by the method of Hesselman (1917*a*). Samples of 100–200 grams, which passed a 2-mm. sieve, are employed. Lunt (1935) and others, however, have used samples of undisturbed soil. The samples, with or without inoculation or addition of nutrient materials, are incubated for a period of several weeks, and the gain or loss of ammoniacal and nitrate nitrogen is then determined. Temperature and moisture are maintained at a favorable level during the incubation period.

The nitrogen requirements of forest trees, in common with requirements for other nutrients, are exceedingly complex and only poorly understood. Requirements evidently vary with species, stage of development, availability of other nutrients, and climatic conditions.

Rebel (1920) indicated that forest stands annually take up about 30-55 lb. of nitrogen per acre and that about 80 per cent of this amount is returned to the soil each year in the leaf fall. The other 20 per cent, about 6-11 lb., remains in the wood of the stems and branches. Values presented by Ramann (1890: 52) are of about the same magnitude. Chandler (1941), working in New York state, found that hardwood tree leaves annually returned to the soil between 14 and 20 lb. of nitrogen per acre. These values are interesting because they indicate in a rough way (1) the quantity of nitrogen absorbed annually, and (2) the substantial annual turnover of nutrient capital.

Data presented by Wilde (1938) suggest that the approximate desirable level of available nitrogen in the upper 8-in. soil layer in forest tree nurseries in the Lake States region is as follows:

Species	Approximate Range in Content of Available Nitrogen, pounds per acre
Jack pine	10-20
Norway pine	20-40
Eastern white pine	30-50
White spruce	40-60

These values of Wilde are based on the combined amounts of nitrate and ammoniacal nitrogen found in forest soils supporting productive stands of the species concerned.

Various aspects of the nitrogen nutrition of forest trees have been investigated by Mitchell (1934, 1939), Gast (1937), and Mitchell and Chandler (1939). Mitchell and Chandler studied the nitrogen requirements of several hardwood forest tree species growing near Ithaca, New York, and Cornwall-on-the-Hudson, New York. By applying different amounts of nitrogenous fertilizers in forest stands and then measuring the growth response and the total nitrogen content of the mature tree leaves, they were able to establish some rather significant relationships between the nitrogen content of the leaves and the needs of the trees.

for soil nitrogen. It appears that the nitrogen content of forest tree leaves after they are mature but before yellowing begins (between August 15 and September 10 in New York state) is, for a given species, proportional to the level of available nitrogen in the soil. The relationship is similar to that shown in Figure 22, page 155. After a curve of this kind has been established for a given species the relative nitrogen level in the soil can be predicted by determining the nitrogen content of the tree foliage. Likewise, a prediction can be made of the extent to which available nitrogen is a factor limiting growth, since the lower is the nitrogen level, the greater is the growth response to an increase in soil nitrogen.

Another relationship established by Mitchell and Chandler is that the nitrogen content of the foliage of different species of trees growing together on the same soil varies, and for a given level of soil nitrogen the foliage of each species possesses a nitrogen content that is more or less characteristic.

These authors obtained significant growth responses to additions of nitrogenous fertilizers in all species investigated. As a result of their work it appears that the species investigated cannot be expected to attain anywhere near their growth potentialities unless the available nitrogen in the soil is equivalent to, or greater than, that evinced by certain minimum concentrations in the leaves. These minimum concentrations are listed in Table 37.

TABLE 37. MINIMUM INTERNAL (LEAF) NITROGEN CONCENTRATION FOR SATISFACTORY GROWTH OF CERTAIN FOREST TREES

(After Mitchell and Chandler, 1939)

Species	Nitrogen Content of Leaves, percentage
<i>Quercus borealis</i>	1.87
<i>Q. alba</i> and <i>Q. montana</i>	2.22
<i>Populus tremuloides</i>	2.00
<i>Acer rubrum</i>	1.70
<i>A. saccharum</i>	1.75
<i>Carya glabra</i>	1.75
<i>Fagus grandifolia</i>	1.95
<i>Nyssa sylvatica</i>	1.83
<i>Fraxinus americana</i>	2.01
<i>Liriodendron tulipifera</i>	2.15
<i>Tilia americana</i>	2.32

Foliar diagnosis, as it relates to the nitrogen nutrition of forest trees, has, in the interests of brevity, been oversimplified in the foregoing treatment. The subject is introduced here to show that in the case of nitrogen, at least, a system of foliar diagnosis, properly conducted, aids considerably in evaluating the nutrient status of a forest soil.

PHOSPHORUS RELATIONS

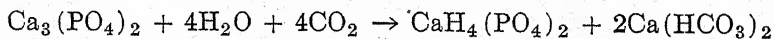
The principal primary source of phosphorus in soils is the mineral apatite, a phosphate of calcium containing a small amount of fluorine or chlorine. Since this mineral is a source of both phosphorus and calcium, it is highly important in forest soils (Tamm, 1920: 297). Phosphorus also occurs in secondary forms as compounds of calcium, magnesium, iron, and aluminum and in organic combinations. In organic form the phosphorus may occur as phytin, phospholipids, and nucleoproteins (Waksman and Starkey, 1931: 173). Truog and Dean (1935) stated that in the majority of soils the inorganic phosphorus exists as calcium or basic iron phosphates. In forest soils with weakly acid or neutral reaction, calcium phosphates may develop; in soils which are strongly leached and poor in calcium, iron and aluminum phosphates predominate (Hackmann, 1930).

The amount of phosphorus in soils is usually small, as the values presented by Pierre (1938) indicate. Because of the great variations in chemical composition, average values of phosphorus content are of little significance. However, as a very rough example, it may be stated that the amount in the surface soil generally varies from less than 0.01 to perhaps 0.20 per cent. Phosphorus is usually more abundant in the surface soil, particularly if considerable organic matter is present, and in the deeper layers than in the intermediate horizons. In the vicinity of Prague, Němec (1929) found that the amount of water-soluble phosphorus in deciduous forest soils decreased rapidly with increasing depth. The highest values were found in the litter layer.

Not only is the total amount of phosphorus in soils apt to be small, but also its availability is generally poor. At any given time the amount of phosphorus in the soil solution is very small; even in fertile soils it is generally less than one-half part per million of solution (Pierre and Parker, 1927). Truog and Dean

(1935) set forth the concepts of readily and slowly soluble phosphorus. Readily soluble phosphorus was regarded as that portion which could be dissolved by a single extraction with a weak solvent, and slowly soluble phosphorus as that which continued to be dissolved during successive extractions. Since form of occurrence greatly influences solubility and hence availability, this characteristic is of greater importance than the absolute amount present. A considerable part of the phosphorus taken up by plants is derived from organic combinations (Gaarder, 1930). Consequently the decomposition cycle of the litter which falls annually is highly important. Hackmann (1930) regarded soils having a mull type of humus layer as generally favorable in their phosphorus relations and soils having a mor type as generally unfavorable. Physical conditions, too, may have an important bearing on the phosphorus supply. Since the roots themselves exert a strong solvent effect on the soil phosphates, it is desirable that physical conditions favor their extensive development. Phosphorus in the form of apatite is only very slowly available. However, Polynov (1937: 157) stated that there are reasons for believing that certain trees, such as pine and larch, may take up the phosphorus of this mineral directly.

Most of the phosphorus salts of cations are only slightly soluble. However, under the influence of water and carbon dioxide the contained phosphorus may slowly become available. The following reaction will serve to illustrate the process:



The insoluble tricalcium phosphate has been transformed to the soluble primary calcium phosphate. Solubility of the calcium phosphate is least at a reaction of more than pH 6.5, and iron phosphates are least soluble at strongly acid reactions, that is, less than pH 4.0. The phosphorus salts of all cations appear most soluble at a pH value between 4.5 and 6.5 (Hackmann, 1930). The ionic form in which phosphorus exists in the soil solution appears to be influenced by the hydrogen-ion concentration. Lyon and Buckman (1943: 36) stated that the PO_4 ion apparently predominates in distinctly alkaline soils. In moderately acid soils the HPO_4 and H_2PO_4 ions predominate whereas at high acidities phosphorus is present largely as H_2PO_4 . It is believed that either an excess or a deficiency of calcium may

interfere with the phosphorus nutrition of plants. Némec (1938) stated that, on soils rich in calcium, fertilization with lime resulted in decreased absorption of phosphorus by spruce. The addition of lime to soils rather poor in calcium resulted in increased absorption of phosphorus. The processes which result in the appearance of phosphorus in the soil solution in forms available to plants are reversible. Phosphorus in available forms may pass into relatively unavailable iron or aluminum combinations, particularly if the soils are highly acid. The fixation of phosphorus in relatively unavailable forms tends to be greatest in soils containing large amounts of active iron and aluminum, for example, in the red and yellow soils of the southeastern United States.

The content of readily soluble phosphorus in soils appears to undergo seasonal fluctuations. Fehér and Frank (1936) and Fehér (1937) have pointed out that in central Europe the citric acid-soluble phosphorus is lowest at the time of maximum vegetation activity, that is, during the summer. Higher values are obtained in the autumn or winter.

The importance of phosphorus from the standpoint of soil productivity has been emphasized by various investigators. Schütze (1869) reported that a relationship existed between the phosphoric acid content of certain sandy soils in Germany and the yield of pine stands. However, in more recent investigations Hennecke (1935) was unable to establish any clear relation between the P_2O_5 content of sandy soils and site quality; other factors evidently were of greater significance. Némec (1929) concluded that the amount of phosphoric acid available in forest soils in the vicinity of Prague was an important factor in forest development. He found less phosphorus in soils supporting coniferous stands than in those under hardwoods. It appeared to Hackmann (1930) that site quality for spruce tended to drop below class II when the soils contained less than 3 mg. of easily soluble (in 1 per cent citric acid) P_2O_5 per 100 grams of dry soil. When the hydrogen-ion concentration was greater than pH 4.0, the same tendency toward poorer site class was noted, even with higher contents of P_2O_5 (up to 7 mg. per 100 grams of soil). He reported that response to fertilization with P_2O_5 was noted only when the soil initially contained less than about 6 mg. of easily soluble P_2O_5 per 100 grams of dry soil.

The problem of the phosphorus requirements of forest trees is extremely complicated. Requirements vary with species, age of trees, and soil conditions. European results indicate that the phosphorus utilized annually for the production of wood in forest stands (rotation about 100 years) amounts to about 0.5–2.5 lb. per acre. This figure represents about 20 per cent of the total annual requirement; the other 80 per cent goes into foliage and is returned to the soil in the leaf fall. For information on the requirements of various species under specified conditions the reports of Mitchell (1934), Kessell and Stoate (1938), and Wilde (1938) should be consulted.

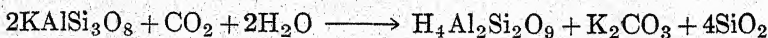
The effects of phosphorus deficiency on various species of tree seedlings have been investigated by Möller (1904), Mitchell (1934), and Jessen (1938). They are in agreement that a deficiency leads to the development of bluish-red, purplish-red, or violet color in the older needles. In some trees the entire needles showed the symptomatic color, but in others only the tips were affected. Phosphorus, like nitrogen, may be deficient in some forest soils.

POTASSIUM RELATIONS

The principal primary sources of potassium in soils are orthoclase, microcline, muscovite, and biotite. Additional sources are the complex secondary aluminosilicates. Most of the potassium in the soil exists in inorganic compounds.

Potassium is usually plentiful, except in sandy soils. The amount in the upper soil generally varies from about 0.15 to 4.0 per cent, thus greatly exceeding the amount of nitrogen and phosphorus. In general, the content of potassium increases with increasing soil depth.

When a "potash problem" exists in soils, it generally concerns slow availability, not actual deficiency, of the element. Under the influence of water and carbon dioxide the minerals containing potassium undergo slow alteration, with the production of simpler compounds which are water-soluble. The alteration may be represented as follows:



Here the mineral microcline alters to potassium carbonate, which is water-soluble, and other products. A small proportion, usually

less than 2 per cent, of the soil potash exists in exchangeable form in the colloidal complex. This, together with the potassium in easily decomposable inorganic and organic compounds, evidently constitutes the principal source of the element for plant nutrition. Wilde and Kopitke (1940) have shown a definite relationship between the amount of available potassium in nursery soils and the cation-exchange capacity; with an increase in exchange capacity there is an increase in available potassium. This fact has an important bearing on the rate of application of soluble potash fertilizers to nursery soils. Heavy applications to soils of low exchange capacity are apt to result in substantial losses by leaching.

There is some evidence that plants may suffer more from potash deficiency in acid soils than in neutral soils (Vítinš, 1924). At the same time it must be pointed out that fertilization with lime on soils very poor in potassium may prove injurious, since the uptake of potassium is inhibited (Süchting, Jessen, and Maurmann, 1937). Némec (1938) also reported that potassium absorption by spruce seedlings is decreased as a result of liming acid soils poor in such nutrients as calcium and phosphorus. The addition of lime to soils rich in nutrients increases the uptake of potassium. To some extent, at least, sodium appears to be able to replace potassium as a plant nutrient.

The amount of citric acid-soluble potassium in soils appears to undergo seasonal fluctuations as a result of biological factors. It is lowest during the summer, when vegetation activity is greatest, and is highest in the autumn or winter (Fehér and Frank, 1936; Fehér, 1937).

As a rule, forest trees use considerably larger amounts of potassium than phosphorus. Henry (1908: 182) stated that the ash of wood removed from the forest commonly contains twice as much potassium as phosphorus. Fehér (1934) has also recognized the relatively high consumption of potash by forest trees. European data indicate that the potassium utilized annually for the production of wood in forests (rotation about 100 years) amounts to about 2.0-8.0 lb. per acre. This figure represents about 35-45 per cent of the total requirement; the other 55-65 per cent goes into foliage and is returned to the soil in the leaf fall. Wilde (1938) has presented data showing the approximate desirable levels of available potassium (as K_2O) in the surface

8 in. of nursery soils in the Lake States. The amounts recommended, based on analyses of productive forest soils, vary from 100 lb. per acre for jack pine to 275 lb. for white spruce. The requirements of plants for potassium, as for other nutrient elements, vary with species, stage of development, and soil and climatic conditions.

CALCIUM RELATIONS

Calcium is of great importance from the standpoint of forest soil fertility. It exerts a strong influence on the physical, chemical, and biological properties of soils (Contejean, 1881; Lang, 1926: 254, 412). Hennecke (1935: 48) regarded calcium, because of its direct and indirect influence on plants, as the most important of the soil nutrients. In addition to its ecological importance calcium exerts a strong influence on soil genesis.

Calcium occurs in both primary and secondary minerals. The principal mineral sources are calcite, dolomite, oligoclase, labradorite, anorthite, augite, hornblende, and gypsum. Although most of the soil calcium exists in inorganic forms, some is held in organic combinations. Substantial amounts exist in exchangeable form in the colloidal complex.

In surface soils the content of calcium usually varies from less than 0.1 to more than 5.0 per cent; generally the amount is less than that of potassium. Soils developed in regions of relatively low rainfall contain larger supplies than soils in humid regions. Hilgard (1892: 24) and others have pointed out that valley soils commonly contain more calcium than do upland soils. The vertical distribution of lime in soil profiles varies considerably. In some soils it is most abundant in the lower horizons, but in others the reverse situation occurs. Trees and other plants which are deep-rooted and have high calcium requirements frequently build up the concentration of lime in the surface soil layers as a result of the contribution made annually by the leaf fall (Taylor *et al.*, 1935: 293; Chandler, 1937; Coile, 1939). The great variability in the calcium content of leaves of different species has already been discussed in Chapter 6.

As a result of the relatively high content of calcium in their

litter, many hardwoods favor accumulation of this element in the surface soil to a greater extent than do most conifers. This fact, which has been demonstrated repeatedly in America as well as in Europe, is indicated by the following data of Alway, Methley, and Younge (1933).

Forest Type	CaO in the Surface
	(Mineral) Soil to a Depth of 2.75 in., percentage
Jack pine	0.22
Jack pine	0.11
Norway pine	0.09
White pine	0.18
Maple-basswood	0.39

A considerable part, perhaps as much as 40 per cent, of the soil calcium is held in replaceable form by the colloidal complex and is readily available. Lyon and Buckman (1943: 27) report that the amount of calcium in exchangeable form in the mineral soils of humid regions is usually six to eight times greater than that of potassium. The hydrolysis and carbonation of various minerals containing calcium, like those containing magnesium, lead to the formation of water-soluble compounds, for example, calcium bicarbonate, $\text{Ca}(\text{HCO}_3)_2$. In this form calcium may be utilized by plants, adsorbed by the soil colloids, or leached out of the soil.

Lack of an adequate supply of calcium in the soil solution is generally the result of deficiency of the element rather than slow availability. Undoubtedly, calcium may be deficient in certain forest soils which are highly acid. It is quite erroneous, however, to assume that calcium is deficient or lacking in all soils having a low pH value. In general, low pH values in mineral soils are symptomatic of low supplies of calcium, but this is not necessarily true of organic matter. In fact, Hoagland and Arnon (1941) suggest that the good growth of some plants in acid organic soils may be related to their high calcium-supplying capacity. Even mineral soils may have a low pH and yet be able to supply large amounts of calcium and magnesium; this is not infrequently true of soils having a high cation-exchange capacity.

Calcium held in exchangeable form is readily replaced by hydrogen ions and has a very high degree of mobility; this fact accounts for its tendency to be easily leached out and lost in the drainage water. It is the most abundant basic radical in fresh-water streams and lakes, representing over 19 per cent of the solid matter held in solution in North American waters (F. W. Clarke, 1924). In soils which support healthy stands of deep-rooted trees a part of the calcium which is moved downward by leaching is absorbed by the roots and subsequently returned to the soil surface in the litter. In such situations losses from the soil body are kept at a minimum. Wiedemann (1925: 87) suggested that this "lime cycle" may be disrupted by clear-cutting, since in the absence of a stand on the ground the lime which would normally be absorbed by roots is washed downward and lost in part, because it is moved below the zone exploited by plant roots or even passes out in the drainage water. This would occur only in regions where the land remained bare after clear-cutting or where the vegetation which developed was extremely shallow-rooted. Losses of calcium in excess of those which are normal and unavoidable occur as a result of fires, soil erosion, and agricultural use of forest soils. Auten (1933) reported that agricultural utilization of forest soils in the Central States has resulted in the loss of a large proportion of the exchangeable calcium and magnesium from the A_1 horizon. Similar conditions have been found in other regions.

Information on the requirements of trees for calcium is scanty and unsatisfactory, but in general they appear to exceed the needs for phosphorus and potash. European data indicate that the amount of calcium utilized annually for wood production in forests (rotation about 100 years) varies from about 5.0 to 25.0 lb. per acre. This figure represents from about 15 to 35 per cent of the total annual requirement, the rest going into the foliage and being returned to the soil in the leaf fall. Chandler (1941) presented data indicating that hardwood stands in New York state return annually to the soil about 65 lb. of calcium an acre. It should be borne in mind that nutrient uptake and nutrient requirement are not necessarily the same and that both vary with species and soil conditions.

Soils having a relatively high capacity to supply calcium are commonly better forest sites than those having small supplies of

available calcium. This fact has been pointed out repeatedly (Hilgard, 1892; Hartmann, 1926; Wiedemann, 1934; Hennecke, 1935: 48). The relationship between calcium content of the soil and site quality is not a strict one. Other factors may be more influential; for example, soils having a favorable water supply usually produce better stands than do dry soils, even though the dry soils may contain higher amounts of calcium and other mineral nutrients.

High concentrations of calcium in the soil may be unfavorable for some species. Dengler (1935: 284) stated that the production of pure stands of spruce is impossible on calcareous soils at low elevations in Germany because of susceptibility to decay. In the warm and dry regions of south Germany, larch shows good development only on soils poor in lime; calcareous soils are unfavorable (Lang, 1933). The regions mentioned are outside the natural range of the species. In this connection it is interesting to note that in the so-called Black Belt in Alabama and Mississippi the calcareous soils (Sumter, Houston, and Bell series) originally supported prairie grasses, not forest. Large amounts of calcium are known to be undesirable in nursery soils. Harmful effects include increased losses of seedlings, due to damping-off fungi, and nutritional disorders. If large amounts of lime are present, iron may be rendered unavailable and the seedlings then become chlorotic. It has been reported that production of acceptable pine seedlings at the Henryville, Indiana, State Nursery became impossible in beds which were surfaced with calcareous sand. The surface 5 in. of soil contained as much as 3 per cent CaCO_3 (Anon., 1938). Most conifers do not require, and will not tolerate, as high concentrations of calcium as most hardwoods.

MAGNESIUM RELATIONS

Magnesium is derived from many minerals, among which may be mentioned biotite, augite, hornblende, olivine, chlorite, talc, serpentine, and dolomite. The content of magnesium in the upper soil layers varies from less than 0.1 to more than 2.5 per cent; very large amounts are sometimes found in soils derived from serpentine. Magnesium, like calcium, may accumulate in the *B* horizon of podzols, but in laterites the amounts present are

very low. The behavior of magnesium and calcium in soils is similar.

The soil magnesium is less readily available than calcium, but more readily available than potassium. It occurs in the form of both inorganic and organic compounds and in replaceable form in the colloidal complex. Hydrolysis and carbonation of the magnesium-containing minerals lead to production of soluble compounds, such as magnesium bicarbonate, $\text{Mg}(\text{HCO}_3)_2$. In this form the element may be absorbed by plants, enter the exchange complex, or be carried away in drainage water.

Magnesium is required for plant growth, but the amounts which are necessary for forest trees are not definitely known. Calculations of the amount annually utilized for wood production in forests (rotation about 100 years) indicate about 1.0–5.5 lb. per acre. This figure appears to be about 25 per cent of the total annual uptake, the other 75 per cent going into the foliage and being returned to the soil in the leaf fall. Soils which contain a large excess of magnesium over calcium have frequently been found infertile (Gordon and Lipman, 1926; Novák, 1937). It has sometimes been assumed that this infertility is due to the toxicity of the magnesium, but this explanation is not wholly satisfactory. Soils with excessive amounts of magnesium usually are poor in other nutrient elements and contain toxic compounds of nickel, chromium, and cobalt. Furthermore, such soils commonly have a high pH value, which may contribute to the nutritional difficulties.

It is probable that magnesium deficiencies may occur, particularly in nursery soils. Möller (1904) reported that a deficiency resulted in the tips of the needles of pine seedlings becoming yellow or orange with the middle portions red and the basal parts remaining green.

SULFUR RELATIONS

The primary mineral sources of soil sulfur are pyrite and gypsum. Sulfur compounds in considerable quantities occur in the atmosphere and are washed down in rains. The amount annually reaching the soil from this source is highly variable, being greatest in regions of industrial development. Perhaps 5–25 lb. per acre annually is a fair average; sometimes the amount is known to exceed 150 lb.

A large part of the soil sulfur appears to be contained in protein molecules in the organic matter. The rest occurs in sulfides, sulfates, and sulfuric acid, and even as free sulfur. Surface soils contain from about 0.03 to 0.4 per cent sulfur (as SO_3). In forest soils a tendency toward accumulation in the upper horizons is frequently observed. This is due to the presence of the element in the organic debris which falls to the ground annually. Heimath (1933) investigated the occurrence of sulfuric acid in forest humus and reported that nearly always the SO_4 -anion content amounted to a substantial part of the titratable acidity.

Although sulfur may be no more plentiful than phosphorus, it is much more readily available. Certain aerobic bacteria can oxidize organic compounds of sulfur, sulfide, and free sulfur to sulfuric acid or sulfates. Sulfate compounds appear to be the source of most of the sulfur taken up by plants. When plant remains containing sulfur are attacked by aerobic microorganisms, most of the sulfur is presumed to be changed to sulfide and subsequently to sulfate. Sulfates are easily leached out of the soil. Under anaerobic conditions sulfates are readily reduced to sulfides, including hydrogen sulfide, which is toxic.

The amount of sulfur utilized annually for wood production in forest stands is probably between 0.1–1.0 lb. per acre. This figure is about 5–25 per cent of the total amount which is removed from the soil; the rest is returned in the leaf fall.

Elemental sulfur and iron, zinc, aluminum, or ammonium sulfates have been used in nurseries to increase soil acidity. Sulfur is oxidized by soil microorganisms to sulfuric acid; compounds such as iron sulfate become hydrolyzed with the formation of the hydroxide and sulfuric acid. The addition of compounds such as ferrous sulfate and ammonium sulfate not only lower the pH of the soil but also supply available iron and nitrogen.

IRON RELATIONS

Iron is an important constituent of many different minerals and rocks; nearly all soils contain an abundance of this element. The amount present in soils varies, depending on the nature of the parent material and the climatic conditions under which the soils developed. In podzols the A horizon is usually poor in iron,

because of removal by leaching, but there is a notable accumulation in the *B* layers. Laterites, which are developed in tropical regions, contain exceedingly large amounts of iron in the upper soil layers. In addition to its occurrence in the form of undecomposed silicates and other primary minerals, iron is found in the colloidal complex. It also occurs in the form of anhydrous and hydrous oxides.

The amount of iron present in soluble form depends on the reaction of the soil and on conditions of oxidation and reduction. Iron is relatively soluble in highly acid soils but is very slowly soluble if the reaction is nearly neutral or alkaline. Under conditions of deficient aeration various insoluble ferric compounds are reduced to soluble ferrous forms, for example, ferrous bicarbonate. Ignatieff (1941) found large amounts of ferrous iron in water-logged forest soils in Canada; very little was found in well-drained soils. When in contact with air, ferrous compounds are easily oxidized. Organic residues in the soil also result in the reduction of iron compounds. McMurtrey and Robinson (1938) stated that the complex ions formed with organic matter hold the iron, even after oxidation, in solution at *pH* concentrations which would precipitate ferric hydroxide.

It is probable that most forest soils contain ample supplies of iron in available form for deep-rooted vegetation, such as trees. In nurseries, however, a deficiency may occur, particularly if the soil has a high *pH*. Seedlings suffering from a lack of iron become chlorotic. Chlorosis, when due to iron deficiency, may be corrected by increasing the soil acidity, adding organic matter, or using an iron salt, such as ferrous sulfate, as a soil amendment or a spray on the plants.

OTHER ELEMENTS OF IMPORTANCE TO PLANTS

Boron, zinc, manganese, and copper in minute amounts are required by plants. The amounts of zinc and manganese absorbed by trees may be considerable. In fact, the exchangeable manganese in forest humus commonly equals and sometimes exceeds in quantity the exchangeable calcium (McMurtrey and Robinson, 1938). Very little is known concerning the importance of these elements in the nutrition of forest trees. All are normally present in soils, at least in minute amounts, and probably seldom limit the development of forest stands.

Aluminum is not believed to be essential for plant development, but it is frequently absorbed in relatively large amounts by trees growing in acid soils. Some investigators have attributed the toxicity of highly acid soils to the presence of active or soluble aluminum.

For many years a livestock disorder called alkali disease has been known in the semiarid great plains of the West. Recently it has been shown that the cause is selenium poisoning. Selenium appears to occur in all soils, but toxic concentrations are for the most part confined to those derived from Cretaceous sediments in the semiarid regions. The selenium is absorbed from the soil by various plants, for example, *Astragalus* spp., and when these are eaten by animals poisoning results. For a thorough discussion of this interesting problem the report by Williams, Lakin, and Byers (1941) should be consulted.

Minor elements of the soil have received increasing attention in recent years, as the rapidly growing volume of literature on the subject indicates. Good discussions of the occurrence of the minor or secondary elements of the soil and their relation to plant development may be found in the reports of McMurtrey and Robinson (1938), Jacks and Scherbatoff (1940), and Scharrer (1941).¹

SOIL ACIDITY

A soil may exhibit one of three reactions, namely, acidity, neutrality, or alkalinity. In view of the fact that most forest soils have an acid reaction, principal attention will be devoted to this condition.

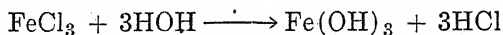
Soil acidity may be caused by the presence of either free organic and inorganic acids or by base unsaturation of the organic and inorganic colloidal complexes. These inorganic colloidal complexes act as colloidal acids. As has been pointed out, the colloid particles may be visualized as being surrounded by a swarm or cloud of ions. The ions surrounding negatively charged colloid particles are cations; those surrounding positively charged particles are anions. In most soils the colloid particles are surrounded by cations. Soils in which the colloids are saturated with

¹ The July and August, 1945, issues of *Soil Science* (60, nos. 1 and 2) are devoted to a symposium on the minor elements.

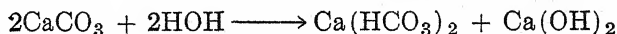
hydrogen give suspensions having a highly acid reaction, whereas base-saturated soils yield either neutral or alkaline suspensions. Forest soils seldom are either wholly hydrogen- or wholly base-saturated and consequently show reaction values intermediate between the two extremes. Leaching is the principal cause of removal of bases and the consequent development of acidity.

Soluble acids, such as carbonic, sulfuric, and nitric, are continually being produced in soils. These acids dissociate in the soil solution, and, if bases are present in the exchange complex, the cationic hydrogen undergoes exchange with a resulting decrease in the acidity of the soil solution. If, on the other hand, the exchange complex is already rather well saturated with hydrogen, the soluble acids tend to remain in solution.

Salts of weak bases and strong acids, for example, FeCl_3 and AlCl_3 , hydrolyze readily and result in an acid reaction:



On the other hand, salts of strong bases and weak acids result in an alkaline reaction:



The proportion in which various ions are present in soil suspensions differs for different soils. In forest soils there is a tendency for the cation hydrogen to predominate because of loss of other cations, such as calcium, magnesium, potassium, and sodium. In soils developed in arid and semiarid regions the content of bases is high, and hydroxyl ions usually predominate over hydrogen ions. Because of this fact most forest soils are acid, and most soils of arid and semiarid regions are nearly neutral or alkaline. An acid reaction, of course, depends on an excess of hydrogen ions, and an alkaline reaction requires an excess of hydroxyl ions. Neutrality depends on equal concentrations of the two.

In pure water, hydrogen ions and hydroxyl ions are present in exceedingly low, but equal, concentrations. Thus, in 10,000,000 liters of pure water at 22°C. there are 1 gram of hydrogen ions and 17 grams of hydroxyl ions. The concentration of hydrogen ions is 1/10,000,000, or, expressed decimally, 0.0000001. It is obvious that expression of low concentrations of ions in terms such as these is awkward. A far more convenient method is to

use pH values, as suggested by the Danish chemist, Sørensen. These values are logarithms of the reciprocal of the hydrogen-ion concentration. For example, the concentration of hydrogen ions in pure water is $1/10,000,000$; the reciprocal is $10,000,000$, and the logarithm of this number is 7. Consequently, the pH of pure water is 7. Solutions having a pH of less than 7 are acid, and of more than 7 are alkaline.

As the concentration of hydrogen ions increases, the concentration of hydroxyl ions must show a corresponding decrease. Conversely, with a decrease in hydrogen-ion concentration, there must be an increase in hydroxyl ions.

It is important to recognize that changes in hydrogen-ion concentration as expressed by pH values are geometric and not arithmetic. Solutions having pH values of 6, 5, 4, and 3, respectively, are 10, 100, 1000, and 10,000 times as acid as water; in other words, they have 10, 100, 1000, and 10,000 times the hydrogen-ion concentration of water. A solution having a pH of 8 is 10 times as alkaline as water, or it has 10 times the hydroxyl-ion concentration of water. As the hydrogen-ion concentration (degree of acidity) increases, the pH value decreases, and vice versa.

TYPES OF SOIL ACIDITY

In strong acids and bases a high proportion of the molecules are dissociated, whereas in weak acids and bases the percentage of dissociation is low. Thus in $0.1\ N\ HCl$ at $18^{\circ}C$. the apparent percentage of dissociation is 92.3, and in $0.1\ N\ CH_3COOH$ it is 1.3. Out of 1000 molecules of the acetic acid only 13 are dissociated; 987 molecules remain unchanged.

Equivalent solutions of acids, for example, $0.1\ N$ hydrochloric and $0.1\ N$ acetic, always contain the same amount of hydrogen, but they may exhibit very different acidic properties, depending on whether they are strongly or weakly dissociated. The strong acids have a high *actual* or *active acidity* and a low *potential acidity*. On the other hand, weak acids have low actual acidity, due to weak dissociation, and high potential acidity, due to the presence of much inactive hydrogen in undissociated acid molecules. Although the actual and potential acidities of weak and strong acids vary, for equivalent concentrations the *titration acidity* is the same. Acidity may be thought of as having two

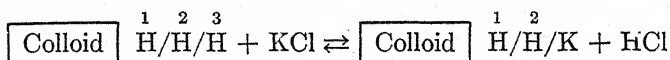
components, namely, an intensity factor and a quantity factor. The intensity factor is the hydrogen-ion concentration (actual or active acidity), and the quantity factor is the total quantity of available acid (titration acidity or normality).

Measurements of soil acidity usually involve the actual or active acidity of soil-water suspensions or water extracts. Two general classes of methods may be recognized: colorimetric and electrometric. Colorimetric methods are commonly used when an accuracy of 0.1–0.2 *pH* is satisfactory. They are simple, inexpensive, and well adapted to field determinations. Electrometric methods are more precise and are in widespread use. Many different electrodes have been employed, but now the glass electrode is very generally used. The soil-water ratio used in preparing soil suspensions or extractions for hydrogen-ion determinations varies with different workers; in general, a ratio of 1–2 is satisfactory for mineral soils, but ratios from 1–3 to about 1–7 have been used for organic materials. Some investigators measure the hydrogen-ion concentration in normal potassium chloride extracts as well as in water. The *pH*-KCl values are generally lower than values obtained in water because of replacement of hydrogen ions in the colloidal complex by potassium ions.

Hydrogen-ion concentration values determined by colorimetric methods may deviate considerably from those determined by electrometric methods. One of the reasons for this difference is the so-called suspension effect, which is particularly noticeable in acid clays. The *pH* of a soil suspension is commonly lower, that is, the acidity is higher, than the *pH* of a soil extract because of the high proportion of hydrogen ions which are held by the colloidal particles. For detailed discussions of the measurement of hydrogen ions the work of Clark (1928) and Snyder (1935) should be consulted.

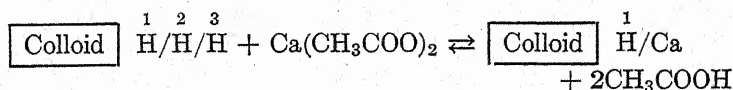
In addition to the dissociated hydrogen ions in soil suspensions (measured in terms of *pH*) there are undissociated molecules of acid and compounds of iron and aluminum which on hydrolysis may form acids. The soil colloids themselves hold hydrogen, together with iron and aluminum ions, which may be replaced by the addition of neutral salts (potassium chloride) or salts which on hydrolysis result in an alkaline solution (calcium or sodium acetate). Neutralization of the acidic constituents may be accomplished by the addition of bases, such as potassium, calcium,

or barium hydroxide. It should be recognized that in addition to the active acidity of forest soils there is a substantial potential or reserve acidity. This potential acidity may be measured by treating a soil with a normal solution of a neutral salt, such as potassium chloride, and titrating the acidity developed in the extract. This is often called the *exchange acidity*. In addition to hydrogen, iron and aluminum in the colloidal complex are replaced by the cation of the neutral salt and appear in the extract. The iron and aluminum chlorides which result hydrolyze, forming hydrochloric acid. Presumably only the rather readily replaceable hydrogen is exchanged, as the following expression indicates:



The hydrogen ion designated as 1 is replaceable with the most difficulty, and the one designated as 3 is replaceable most easily. As a result of the exchange acidity a KCl extract of an acid soil is much more acid than a water extract.

If a normal solution of calcium acetate is used instead of a solution of potassium chloride, the titratable acidity developed in the extract is called the *hydrolytic acidity*. More of the exchangeable hydrogen of the colloidal complex is replaced, as is indicated in the following reaction:



The greater exchange of hydrogen results from the greater replacement capacity of the divalent calcium and the formation of weakly dissociated acetic acid in the buffered acetate solution. Even when the active acidity and exchange acidity are low, there is a substantial amount of hydrolytic acidity, which is regarded as the weakest form of acidity in soils (Potapov, 1933). This value is seldom employed by American soil scientists, but its use in Europe is not uncommon.

Titration curves have been found helpful in characterizing the acid constituents of soils. Information of a quantitative as well as a qualitative nature is obtained. The neutralization curves obtained when organic and inorganic colloids are titrated, using bases such as sodium or barium hydroxide, resemble those of weak acids. Sometimes no break seems to occur in the titration curve,

but often one occurs between pH 6 and 8 and may be regarded as indicative of a practical end point. Neutralization curves of organic and inorganic colloids differ materially. Feustel (1939) reported neutralization values (the amount of alkali required to produce pH 7) of 3.1–5.0 m.e. per gram for humic acids and only 0.23–0.57 for inorganic colloids. Of course, pronounced differences are exhibited by different colloids within each of the two broad groups. Anderson and Byers (1936) found the colloids of chernozem soils to have much higher neutralization values than colloids of laterite soils. Values for colloids of prairie and gray-brown podzolic soils are intermediate between those of chernozem and laterite soils. In podzol soils the colloids differ very materially in the various horizons. Except in podzols, there seemed to be a relationship between the general level of fertility and the amount of alkali required to produce pH 7 (the cation-exchange capacity of the colloid at neutrality) in the soils investigated by Anderson and Byers. These authors expressed the view that, when this value for the colloid is more than about 0.45 m.e. per gram, the crop-producing capacity is apt to be limited more by water supply than by lack of chemical constituents.

FACTORS AFFECTING THE HYDROGEN-ION CONCENTRATION OF SOILS

The hydrogen-ion concentration of soils is influenced by many factors, among which the following appear most important: time of sampling, drying of soil, kind of soil, vegetation, and silvicultural practices.

EFFECT OF TIME OF SAMPLING

The pH of a forest soil is not constant but shows variations during the course of a year. G. R. Clarke (1924) stated that in England the pH was highest during the period of most active tree growth. On the other hand, Nehring (1934), working in Germany, recorded the highest pH values in the winter and the lowest in the summer. He found that the variations generally did not exceed 0.8 pH. Joffe (1935) reported that pH values in forest humus layers were highest in the autumn. He explained this maximum as the result of bases which were released from the freshly fallen leaves. Seasonal variations in the pH of soils are

probably related to climatic conditions and to the nature of the vegetation. These variations need to be considered when comparing pH values obtained at different times.

EFFECT OF DRYING OF SOIL

On the basis of numerous investigations, of which that by Bailey (1931) is particularly important, it seems clear that drying of soils is, as a rule, accompanied by a change in reaction, generally in the direction of increased acidity. The amount of change is usually less than 0.10 or 0.20 pH . The use of air-dried soils for the determination of pH is general and seems justified because it affords a comparable basis. However, in critical studies of soil reaction it may be desirable to use fresh soil samples.

EFFECT OF KIND OF SOIL

Soils representative of the great groups are known to differ materially in acidity. Bailey (1933) found that podzol soils were most acid, followed in order of decreasing acidity by lateritic soils, podzolic soils, humid prairie soils, chernozem soils, and desert soils. The surface layers of chernozem soils were slightly acid to neutral, and all the desert soils were alkaline.

Marked differences are frequently observed in the pH of different horizons in a given soil body. Thus in podzols the H layer or the A horizon is commonly most acid; that is, these layers have the lowest pH . Maximum pH values are commonly encountered in the lower part of the B horizon (Aaltonen, 1937). In the gray-brown podzolic and brown forest soils the A_1 horizon is generally most acid. Two tendencies are noted in profiles of laterites: (1) the soil becomes more acid with increasing depth, or (2) the soil maintains the same pH throughout the profile.

Altitude, aspect, and drainage influence the pH of soils. In general, pH values decrease with increasing elevation, and in the northern hemisphere the soils of north slopes tend to be more acid than those of south and west slopes. Under conditions of poor drainage highly acid conditions are apt to develop. Holmes, Hearn, and Byers (1938) investigated soils of the Norfolk family

in North Carolina and found the lowest pH values in those having the poorest drainage.

EFFECT OF VEGETATION

Vegetation exerts a strong influence on soil acidity through the litter which it supplies and through its effect on soil temperature and moisture. In general, soils supporting conifers tend to be more acid than those supporting hardwoods (Hesselman, 1926; Němec and Kvapil, 1926; Alway, Methley, and Younge, 1933; and many others). This situation is to be expected in view of the relatively high acidity and low basic buffer content of most conifer leaves, as compared with most hardwood leaves. Species whose foliage contains a high content of bases tend to prevent the development of excessive acidity in the surface soil layers. In fact, instances are known where the surface soil layers are less acid than the deeper layers because of this concentration of bases in the litter (Doyle, 1935).

The introduction of hardwoods in conifer stands is sometimes known to result in decreased acidity. The effect of this measure depends on the hardwood species employed and on the nature of the soil.

It is generally recognized that soils supporting grasses are less acid than those supporting forest. When trees invade grassland, an increase in soil acidity usually occurs.

EFFECT OF SILVICULTURAL PRACTICES

Silvicultural operations, such as heavy thinnings and reproduction cuttings, are known to result in decreased soil acidity in some instances (Olsen, 1923; Hesselman, 1926; Wiedemann, 1934, 1937). Fertile soils respond more readily than infertile soils. The reduction in acidity appears to result from higher soil temperature, more precipitation reaching the soil, and changes in the soil flora and fauna. In excessively dense stands soil acidity is apt to be higher than in more open stands. Fires generally result in a reduction of soil acidity because of ashing of part of the organic matter and possibly because of decomposition of soil minerals.

ECOLOGICAL IMPORTANCE OF SOIL ACIDITY

There is very little evidence that low pH *per se* is responsible for poor growth of forest trees. Even in agricultural plants, which presumably may be more sensitive to high concentrations of hydrogen ions than are trees, profoundly adverse effects of hydrogen ions as such seem to occur only at exceedingly low pH values (pH 3 or less). Investigators have repeatedly found that low pH of forest soils is not incompatible with high site quality (Hesselman, 1926; Wiedemann, 1932). In northern Europe soils having pH values of 4.0 or less are known to be highly productive, even for hardwoods, such as beech. Wiedemann (1937) reported that many hundreds of measurements of the degree of acidity in spruce soils of Germany failed to give any clear indication of site quality.

Although the direct effects of high soil acidity appear unimportant, the same statement cannot be made of the indirect effects. As a consequence of a high concentration of hydrogen ions the soil fauna and microflora may undergo change in an unfavorable direction, the physical properties may be impaired, and nutritional disturbances due to deficiencies of nutrients (calcium or magnesium) and to toxicity of certain elements (manganese or aluminum) may develop. Just how important these considerations are in the development of forest stands cannot now be stated. It seems probable, however, that seedlings are more sensitive than older trees.

Alkaline soils do not seem to be as favorable for the development of forest vegetation as are acid soils. Years ago Ruffin (1855) pointed out that calcareous soils usually support grasses. It may be argued that the general absence of commercial forests on alkaline soils is due to climatic aridity as such and not to the nature of the soil. Frequently this is true, but the existence of natural prairies on calcareous soils, for example, the Black Belt in Alabama and Mississippi, in forest regions demonstrates that soil alkalinity is more favorable for grasses than trees. The unfavorable effects of soil alkalinity may be due not alone to the toxicity of the hydroxyl ions but also to the unavailability of nutrients such as phosphorus, iron, or manganese.

In general, tree seedlings develop best in soils having reaction values between pH 4.5 and 6.0. With reaction values of less than about pH 4.5 unfavorable effects such as those just mentioned

may appear, and with reaction values higher than about pH 6.5 damping-off fungi, eel worms, and other organisms may result in serious seedling losses (Wilde, 1934). Losses of pine, spruce, and Douglas fir seedlings in forest-tree nurseries as a result of damping-off are reduced if the soils are maintained in a fairly acid condition (Hartley, 1930).

Sometimes it has been necessary to acidify nursery soils in order to produce satisfactory coniferous stock. Materials such as ammonium sulfate, aluminum sulfate, ferrous sulfate, sulfuric acid, acetic acid, and acid humus have been used. As a rule, hardwoods develop best in soils somewhat less acid than those in which conifers attain their optimum growth.

Rather narrow pH limits are sometimes specified for the optimum development of various species. In view of the fact that the influence of pH varies in different soils any specification of narrow reaction requirements becomes questionable. Furthermore, it should be recalled that pH varies in different soil horizons and that seasonal differences occur within given horizons. Trees grow normally within rather wide variations of pH . The most useful list of the pH preferences of plants is that of Spurway (1941).

BUFFER ACTION

Soils exhibit a distinct resistance to change of pH on addition of acids or bases. This opposition to change of reaction is called *buffering*. In view of the fact that soil colloids contain both acidoid and basoid constituents it is not surprising that they are buffered against both acids and alkalies.

Buffering results from the presence of very weak acids or bases, or salts of weak acids and strong bases, or salts of weak bases and strong acids. As an illustration of buffering against alkali the behavior of an acetic acid solution may be mentioned. Acetic acid, like other weak acids, is only slightly dissociated, with the result that most of the hydrogen occurs in molecular combination. If sufficient base is added to the solution to neutralize the dissociated hydrogen ions, more of the acetic acid molecules ionize and little change in pH is observed. Buffering against acid may be illustrated by the behavior of a solution containing a salt of a weak acid and strong base, such as sodium acetate. This salt

ionizes, forming CH_3COO^- and Na^+ ions; and, when an acid such as HCl is added, the hydrogen ions associate with the acetate ions forming acetic acid. Because of the incomplete dissociation of the acetic acid formed, most of the hydrogen ions added in the HCl are inactivated, and little change in pH occurs.

Soil colloids behave in much the same manner. The addition of a base to colloidal material results in replacement of hydrogen ions by the added cation; the liberated hydrogen ions then unite with the added hydroxyl ions, forming water. On the other hand, when an acid is added, the hydrogen ions replace cations of the colloidal complex (Ca , Mg , K , Na), which in turn form neutral or nearly neutral salts. In either case little change in pH is noted in well-buffered soils.

The buffering capacity of soils and plant materials against acids has received considerable attention in recent years. Relatively high buffer capacity of forest soils is regarded as favorable. Highly siliceous sands and the A_2 horizon of podzol soils are poorly buffered.

OXIDATION-REDUCTION POTENTIALS IN SOILS

One of the fundamental features of an oxidation-reduction process is an exchange of electrons. Oxidation occurs when an ion loses electrons, and reduction occurs when electrons are gained. By measuring the oxidation-reduction potential or electron pressure, the relative state of oxidation in soils may be ascertained.

Only in recent years have oxidation-reduction potentials of soils been investigated. The E_h , or oxidation-reduction potential, values vary, depending on the methods used; these methods are still in an experimental stage. However, promising results were obtained in an investigation of the relations of oxidation-reduction potentials to yield of apple trees by Bradfield *et al.* (1934). Relatively high E_h values of the soil were associated with high yields, and low values with low yields. Future investigations may demonstrate that measurements of this type are useful in the evaluation of forest soils.

For discussion of oxidation-reduction potentials and their measurement Clark (1928), Bradfield *et al.* (1934), Volk (1939), and Dirks (1940) should be consulted.

FERTILIZATION OF FOREST AND NURSERY SOILS

Few soils are so fertile that a response is not obtained from the addition of some fertilizer constituent or soil amendment. However, fertilization cannot be justified by the mere fact that crop yields are increased by a measurable amount. For economy the benefits derived from fertilization must exceed the costs involved.

At the present time fertilization of forest soils cannot be justified in America. Even in Europe, where costs are relatively low, such fertilization is still largely on an experimental basis. For a detailed discussion of the results of fertilization in German forests the report by Wiedemann (1932) should be consulted.

In nursery soils a quite different situation exists. Here the drain on fertility elements is very great because of the high density of the seedlings (often 50 or more per square foot) and the fact that the entire crop, both tops and roots, is periodically removed. Furthermore, the young trees have rather high requirements for nutrient elements. It is now generally recognized that fertilization is necessary in permanent nurseries.

No one method of fertilization can be applied to all nurseries with equal success. Like silvicultural practice in general, nursery soil fertilization is a distinctly local proposition. The nature of the soil, climatic conditions, and species grown all have a bearing on the problem. The best way to determine proper practice in a given nursery is to experiment on a small scale with different kinds and amounts of fertilizers. Physical and chemical analyses of the soil, pot culture tests, and careful observation of the seedlings for nutrient deficiency symptoms all contribute information which aids intelligent experimentation.

For details of nursery soil fertilization Wahlenberg (1930), Toumey and Korstian (1931), Stoeckeler (1933), Wilde (1937, 1938), Wilde and Hull (1937), Lunt (1938), Mitchell (1939), Wilde and Patzer (1940), and Auten (1945) should be consulted.

INFLUENCE OF FIRES ON THE CHEMICAL PROPERTIES OF SOILS

When organic matter is destroyed by fire, the nonvolatile constituents remain on the mineral soil surface in the form of ash. The immediate effect is to alter the reaction of the soil in the direction of alkalinity and to render mobile substantial amounts of calcium, magnesium, potash, and phosphorus. Rains move

the soluble material into the soil, where some is held in the upper horizons by the colloidal complex, part moves so deeply that it is below the reach of plant roots, and part moves into the drainage water and is entirely lost. If the soil has a relatively high cation-exchange capacity and if a cover of vegetation develops promptly, losses by leaching are probably small; otherwise they are large. Losses of nutrient elements as a result of fires are greatest in areas of coarse sandy soils. Erosion after fires may result in loss of both ash material and fertile top soil with consequent depletion of the nutrient capital of the site. In addition to rendering quickly available the nutrients contained in the organic matter, fires probably result in a release of elements contained in the soil minerals (Ehrenberg, 1922: 454).

The influence of fires on the nitrogen relations of soils has already been considered and will be only briefly mentioned here. Substantial losses of soil nitrogen occur as a result of fires, but in the longleaf pine soils of the South these losses subsequently may be more than offset by increased fixation of atmospheric nitrogen by free-living and symbiotic organisms. In northern forests having mor types of humus layer the total amount of nitrogen may be reduced as a result of fires, but the available amount appears to be increased.

The effects of fire are extremely disastrous in situations where almost purely organic soils rest on rocks or on very thin layers of loose mineral material. In such areas the soil itself, being composed of carbon compounds, is destroyed. Certainly hundreds, and perhaps thousands, of years must elapse before forest vegetation can redevelop on situations like that shown in Figure 23, page 183.

Examples where the use of fire can be justified solely on the basis of a favorable effect on the chemical properties of a forest soil are rare. In regions or types where fire is used or advocated as a silvicultural tool the primary considerations usually involve reduction of competing vegetation, decrease of fire hazard, sanitation benefits, exposure of a mineral soil seed-bed, or improvement of forage, and not to the soil factor as such. The great majority of fires must be regarded as harmful to the soil. The degree of injury varies with the nature of the soil, the intensity and frequency of the fires, and the rapidity with which a new cover of vegetation develops.

MAINTENANCE OF SOIL FERTILITY

Maintenance of soil fertility is a responsibility which rests directly on the shoulders of the silviculturist. He should recognize that the soil capital with which he works may be depleted, maintained, or perhaps increased, depending on the way it is handled. In forestry, as in agriculture, a fertile soil is essential to maximum production.

The practical measures which may be taken to conserve and improve forest soil fertility are mostly indirect. They include choice of species adapted to the site, control of stand density, avoidance of prolonged exposure of bare soil, and protection against injurious agencies, such as fire, grazing, and erosion. As a general rule, deep-rooted species, because of their ability to utilize nutrients in the lower soil layers, conserve soil fertility better than shallow-rooted species. Likewise, species which absorb relatively large amounts of bases contribute more to maintenance of a high nutrient level in the surface soil layers than do those having lower requirements. Mixtures are commonly believed to exert a more favorable influence on the soil than pure stands. This fact has led some individuals to the unfortunate generalization that pure stands and poor soil conditions are practically synonymous. German experience is commonly believed to substantiate this view. As a matter of fact, unbiased examination of the European literature leads to the conclusion that any such generalization is untenable.

There is increasing recognition among foresters that excessive stand density may lead to unfavorable soil conditions. The principal direct effects seem to be lowered soil temperature and decreased rainfall reaching the ground. Indirect effects are an increase in soil acidity and unfavorable changes in soil fauna and flora; these in turn influence the physical and chemical soil properties.

APPRAISAL OF NUTRIENT AVAILABILITY

No simple techniques are known for definitely ascertaining either the amount of soil nutrients available to forest trees or nutrient deficiencies. Interrelations between plants and the media in which they grow are so complex that any simple solution of the problem is precluded (Hoagland and Arnon, 1941).

Two general approaches have been made: (1) chemical tests or analyses, and (2) biological tests. Chemical analyses may show total quantities of the various constituents present, or they may indicate only the amounts which are readily soluble or available. Total analyses are of little value in determining amounts of available nutrients. Partial analyses indicating available amounts of the various elements are more useful, but even with this type of information it is impossible to arrive at a definite conclusion as to actual nutrient availability to trees growing in the forest. The highly empirical nature of chemical tests of available nutrients should be recognized.

In spite of their limitations the approximate quantitative estimations of available nutrients afforded by chemical tests are of interest to the silviculturist from several points of view. The information supplied contributes to an understanding of the influence of species composition, silvicultural practices, and other agencies on the chemical character of the soil. It may in time aid materially in the classification of soils for the production of forest crops. Interpretation of chemical data in terms of forest-tree growth requires the type of knowledge which has been acquired by agronomists in field tests of yields of specific crops on specific soils. The fact that the surface 6 in. of a given forest soil contains 50 lb. of available P_2O_5 per acre means little until it is correlated with growth or yield of specific species. Very little is at present known about the relation between definite soil fertility levels and site quality of forest soils. About all that can be said is that, in general, production increases with greater amounts of available nutrients and that the level of fertility required for satisfactory growth of different species varies. Relatively richer soils are required for hardwoods than for conifers. It seems clear that the results of chemical analyses can be interpreted only in the light of a background of considerable local experience.

Methods for chemical analyses of soils may be found in publications such as that of Spurway (1938), Morgan (1941), Wright (1939), the Association of Official Agricultural Chemists (1940), Anderson *et al.* (1942), Piper (1944), and Peech and English (1944).²

² The January, 1945, issue of *Soil Science* (59, No. 1) is devoted exclusively to methods in chemical analyses of soils.

Conductivity measurements of soil extracts or soil-water suspensions are believed worthy of consideration in connection with other soil tests. Electrical methods for determining the soluble-salt content of a soil are based on the fact that the electric current is conducted by the salt in solution and that the conductance of the solution is determined largely by its concentration. Conductivity increases, or resistance decreases, as the salts in solution increase. Only total concentrations are indicated; no information is supplied concerning the relative quantities of different ions in solution. Details concerning the use of the electrolytic bridge for quantitative determinations of the salt content of soils may be obtained from Davis and Bryan (1910) and Wright (1939).

Biological tests in which plants are used to evaluate nutrient availability have advantages over chemical analyses of the soil. The fertility of the soil may be judged simply by the growth or dry weight of the plants produced either in pot cultures or in field trials, or the whole plants or their leaves may be analyzed for content of specific elements. Analysis of leaves, commonly referred to as foliar diagnosis, has received considerable attention in recent years, notably by Mitchell and Chandler (1939). Promising results have been obtained, and the method will undoubtedly prove a useful aid in solving the exceedingly complex problem of evaluating soil fertility. It should be emphasized, however, that interpretation of leaf analysis data is not simple; climatic, physiological, and soil factors have to be considered (Hoagland and Arnon, 1941).

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Chapter 11

SOIL FORMATION

As a result of the action of natural forces and living organisms, rocks and minerals exposed at the surface of the earth are transformed in the course of time into soil. The rate of transformation and the character of the soil which ultimately results vary with the nature of the environment and the parent rocks and minerals. Consequently, very different soils are encountered in differing climatic regions; even in regions which are climatically uniform, soils may differ greatly as a result of variations in surface geology.

As soil develops from the parent material, decomposition of minerals continues, organic matter is accumulated, and materials in suspension and solution are translocated in the soil body and in part removed in the drainage waters. New chemical compounds are formed, and soil horizons are developed. These processes continue until a condition of near-equilibrium with the environment is attained; the soil body is then mature, and the acquired characteristics dominate the inherited ones. The state of soil development referred to as maturity should be regarded not as static but as somewhat dynamic.

Young, imperfectly developed soils are immature. Immaturity may be due to the fact that the soil-forming factors have been operative for only a relatively short period of time. Poor drainage, parent material containing large amounts of bases, and erosion are other factors which inhibit development of soils to maturity.

Three categories of soils are commonly recognized, namely, zonal, intrazonal, and azonal. Zonal soils are those having well-developed characteristics which reflect the influence of the active factors of soil genesis—climate and living organisms. Intrazonal soils have more or less well-defined characteristics which reflect the dominating influence of some local factor, such as relief, parent material, or age, over the normal effect of climate

and vegetation. Azonal soils lack well-developed profile characteristics; this situation results from the youth of the soil material or from peculiar conditions of parent material or relief.

THE FACTORS OF SOIL FORMATION

Soil formation is conditioned by a number of factors, among which parent material, climate, living organisms (plant and animal), relief, and time are most significant. These factors have received detailed consideration by Neustruev (1927), Byers *et al.* (1938), and Jenny (1941).

PARENT MATERIAL

Parent material usually refers to the unconsolidated mass from which the soil profile develops, not to the bedrock itself. On weathering, rocks give rise to parent material; this may well be regarded as the first step in soil development. Since various soil-forming minerals and rocks and the processes leading to their disintegration and decomposition have been considered in earlier chapters, they need not be discussed here.

The nature of the parent material has a bearing on the type of soil resulting and on the rate of profile development. Both the texture and the content of nutrients in soils are influenced by the nature of the parent material. In the coastal plain region of New Jersey the relatively fine-textured and fertile Sassafras loam soil has developed from mixed clays, silts, and gravels, whereas the relatively coarse-textured and infertile Lakewood sand has developed from quartz sands. Profile development is more rapid in materials that are easily permeable to water and contain low amounts of bases than in materials less readily permeable to water and containing large amounts of bases, especially lime. Brown forest soils or brown podzolic soils may occur adjacent to podzols, the brown forest or brown podzolic soils having developed on heavy-textured glacial till and the podzols on light-textured siliceous sands. The presence of calcium carbonate in soils of humid regions is known to impede development of the zonal soil. Highly calcareous materials frequently give rise to immature soils called rendzinas.

By way of summary it may be stated that parent material influences both the physical and the chemical properties of soils.

The influence is so profound that within a given climatic region different azonal, intrazonal, and zonal soils may be developed. Much of the local variability of soils is due to variations in the parent material from which they were derived. It should be borne in mind that the differentiating effects of parent material are greatest in immature soils and that many forest soils are immature.

CLIMATE

The important role of climate in the process of soil formation was recognized half a century ago by the American Hilgard and the Russians Sibirtsev and Dokuchaiev. Hilgard (1892) expressed the situation in these words: "Since soils are the residual product of the action of meteorological agencies upon rocks, it is obvious that there must exist a more or less intimate relation between the soils of a region and the climatic conditions that prevail, or have prevailed therein." Climate influences soil

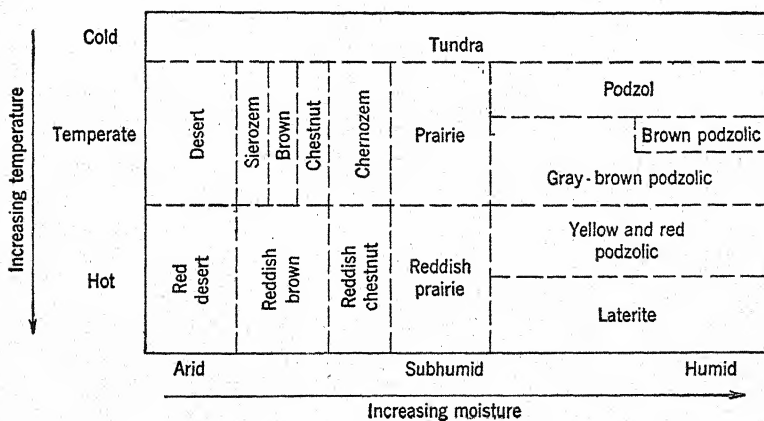


FIG. 37. The relationship of the principal zonal soil groups to climate.

formation directly, because it affects rock weathering, and indirectly, because it affects plant and animal life. The geographical distribution of the zonal soils of the world is chiefly determined by climate and living organisms, principally vegetation (Figure 37).

Moisture relations are in part a function of climate. They

influence soil genesis to a remarkable degree. Thus soils developed in humid regions are subjected to intense leaching and present profiles wholly different from those of soils developed in arid regions. In Table 38 are presented data which illustrate this fact.

TABLE 38. AVERAGE CHEMICAL COMPOSITION (PERCENT-AGE) OF SOILS OF ARID, PRAIRIE, AND TIMBERED HUMID REGIONS

(After Coffey, 1912)

Region	Number of Samples	CaO	MgO	K ₂ O	P ₂ O ₅
Arid	318	2.65	1.20	0.71	0.21
Prairie	215	1.09	0.51	0.43	0.18
Timbered humid	743	0.41	0.37	0.37	0.16

The amount of water which percolates through the soil in a given region depends on various climatic factors, among which are the amount and intensity of precipitation, atmospheric humidity, temperature, and length of the frost-free period. Total annual precipitation alone is not a good criterion of the effectiveness of water in soil formation. Long-continued rains of low intensity are more effective than rains of short duration and high intensity. Precipitation of the brief, intense type is liable to result in excessive run-off and erosion. Low atmospheric humidity, high temperature, and a short frost-free period all tend to decrease the amount of water which percolates through the soil.

In general, it appears that with more rainfall the content of soil clay increases. This change is to be expected as a result of increased weathering with greater precipitation. With increasing rainfall there also is an increase in hydrogen-ion concentration, exchangeable hydrogen, and depth of the carbonate horizon resulting from leaching of bases. Jenny (1941: 118) indicated that the nitrogen and organic-matter of the surface soil becomes higher as moisture increases.

Temperature influences soil formation through its effect on the rate of weathering of rocks and soil materials. As a rule, the rapidity of chemical reactions approximately doubles for each rise of 18°F. in temperature. Soil-forming processes are at a minimum in regions of low temperature and at a maximum in humid regions having high temperatures.

With increasing temperature the depth of weathering and the content of clay increase. On the other hand, the content of both nitrogen and organic matter in soils decreases from regions of low temperature to regions of higher temperature.

The role of climate in soil formation has been investigated by means of various expressions or criteria which are based on combinations of factors. Meyer (1926) employed values which he called *N-S* quotients, derived by dividing the precipitation (*Niederschlag*) in millimeters by the absolute saturation deficit (*Sättigungsdefizit*) in millimeters of mercury, calculated for the full year, the frost-free period, or individual months. High *N-S* quotients signify humid climates. Rain factors, obtained by dividing the precipitation in millimeters by the temperature in degrees Centigrade, have also been widely employed in studies of the relation of climate to soils. More recently Thornthwaite (1931) has developed a classification of climates based on "precipitation effectiveness" and "temperature efficiency." In a general way the climatic types of Thornthwaite correspond to many of the zonal soils.

Close and invariable relationships between climatic zones and soil zones should not be expected, because climate is only one of the factors responsible for soil development.

LIVING ORGANISMS

Plants influence soil development by their alteration of the microclimate, by their residues which rest on or are incorporated in the mineral soil body, and by their transfer of elements from the lower to upper horizons. In addition, vegetation favors soil development by diminishing erosion. Taber (1943) has pointed out that in the interior of Alaska the dense cover of vegetation retains rock waste on the valley slopes until it is disintegrated by frost action.

The effect of different types of vegetation on soil development varies, as would be expected in view of the great differences in life-form, depth of rooting, and physiological behavior. A substantial part of the organic remains contributed to the soil by annuals consists of roots which remain in the soil; most of the organic matter which is annually supplied by forest trees drops on the mineral soil surface. This fact explains in part the marked differences of organic matter content which occur in the

upper mineral soil layers of grassland and forest soils. Deep-rooted species contribute more to soil development than species having shallow roots. The deep roots aid in weathering of the parent material in the lower layers and absorb nutrient elements which are subsequently returned to the soil surface in the leaf fall. Species vary greatly in the amounts of chemical elements absorbed from the soil. As was pointed out in Chapter 6, certain species, such as white ash, the hickories, basswood, and tulip tree, absorb larger amounts of bases than do most conifers. It should be recognized that in humid regions the movement of soluble materials by percolating water is continuous and in one direction only, that is, downward. Gradual impoverishment of the surface soil is the inevitable result of this process. Vegetation plays an important role in partially counteracting the effects of leaching. As a broad generalization, it may be stated that hardwoods are usually more efficient in this respect than are conifers.

Microorganisms aid in rock weathering and in the decomposition of organic matter. They may well be regarded as indispensable in soil development. Many of the chemical processes in soils are traceable directly or indirectly to the activities of the soil fungi and bacteria. The higher animals also contribute to soil formation by mixing inorganic and organic materials, by loosening the soil, and by aiding in the decomposition of organic matter.

The relationship between soils and living organisms is close. Each influences, and in turn is influenced by, the other. It is a well-known fact that replacement of forest vegetation by grasses, or vice versa, leads to profound changes in soil conditions. In the same way replacement of a hardwood forest by a coniferous stand, or of a coniferous stand by a hardwood forest, leads to soil changes. This simple fact should be recognized by the silviculturist. He should appreciate the practical impossibility and should question the desirability of maintaining under diverse forest types a common soil condition.

RELIEF

Relief influences soil development through its effect on water relations, erosion, temperature relations, and vegetation cover.

Rain falling on soil tends to run off steep slopes and collect in depressions. The result is that soils on steep slopes receive less water and soils in depressions receive more water than those of flat areas. Soil development on steep slopes tends to be inhibited by the reduction of the amount of water entering the soil. Most forest soils, particularly in the region of podzols, are characterized by microrelief features resulting from the uprooting of trees or from other causes. It has frequently been observed that podzolization proceeds more rapidly in depressions, which are moist, than on adjacent elevated situations, which are dry. On the other hand, if the amount of water in the soil is excessive because of poor drainage, intrazonal soils such as the ground-water podzol or half-bog may develop. So-called normal soils are usually those which have developed on gently undulating, but not strictly level, well-drained upland areas. On flat areas in humid regions soils finally reach a condition referred to as senility; the surface horizons are leached, and the *B* horizons are claypan or hardpan. Intrazonal soils of this kind are planosols.

Rapid normal erosion is unfavorable from the standpoint of soil formation because the soil is maintained in a youthful stage of development. This situation exists in many forest soils on steep slopes. On the other hand, extremely slow erosion, such as occurs on flat areas, is likewise unfavorable. For development of so-called normal or mature soils, erosion should keep pace with the rate of soil development but should not exceed it.

In the northern hemisphere, south and west slopes are warmer than north and east slopes. This fact influences weathering and also frequently causes differences in composition and density of the vegetation. Red podzolic soils frequently extend to higher altitudes on south slopes than on north slopes in the southern Appalachians. In the West, podzol soils are more common and tend to occur at lower elevations on north slopes than on south slopes. The influence of exposure on soil development is particularly noticeable in regions having relatively low rainfall.

Swederski (1933) investigated the relationship between mountain soils and relief in the east Carpathians. He reported that the influence of relief was expressed chiefly in the depth of soil, its mechanical composition, and its content of humus. Depth of soil was least on steep slopes.

TIME

The length of time required for soil development depends on the factors previously considered, namely, parent material, climatic conditions, living organisms, and relief.

It is obvious that the time required for development of a soil body in near-equilibrium with its environment depends, first of all, on the initial state of the substratum. A longer time is required for the development of soil starting from bedrock than is necessary if parent material is available. It is probable that normally more time is required for accumulation of the parent material than is needed for subsequent development of the soil profile. Bedrock in New England has in many places weathered so slightly since the Wisconsin stage of glaciation that glacial striae are still clearly evident. On the other hand, it appears that immature podzol soils may have developed from dune sands on Cape Cod in about 100-200 years. Tamm (1920) reported that after drainage of Lake Ragunda in Sweden in 1796 perceptible podzolization occurred in 100 years under forest. However, at least 1000-1500 years was considered necessary for development of a leached A_2 horizon 4-6 in. thick. Soil profile development in humid regions proceeds more rapidly in coarse-textured, acid parent materials than in fine-textured, basic materials.

Less time is required for soil development in humid, warm regions with luxuriant vegetation than in dry or cold regions with scanty vegetation. The reasons are obvious from the preceding discussion.

Relief has an important bearing on the age of soils through its influence on erosion and deposition. Situations where either erosion or deposition is active have youthful soils.

It should be clearly recognized by the silviculturist that soil immaturity is in no way incompatible with high site quality. Young soils are frequently more fertile and support better forest stands than do mature soils in the same region.

PROCESSES OF SOIL FORMATION

DEVELOPMENT OF ZONAL AND INTRAZONAL SOILS

The soil-forming processes responsible for development of the major zonal and intrazonal soil groups are: (1) podzolization, (2) laterization, (3) formation of peat and poorly drained soils

AFTER GLINKA, MARBUT AND OTHERS.

TUNDRA.

PODZOLS.

GRAY-BROWN PODZOLIC SOILS.
BROWN FOREST SOILS, ETC.

LATERITIC SOILS (LATERITES,
TERRA ROSSA, RED LOAMS,
RED AND YELLOW SOILS.)

PRAIRIE SOILS AND
DEGRADED CHERNOZEM.

CHERNOZEMS.

CHESTNUT AND
BROWN SOILS.

SIEROZEMS AND DESERT SOILS.

SOILS OF THE MOUNTAINS AND
MOUNTAIN VALLEYS (COMPLEX).

A — SOIL MATERIAL LARGELY OF ALLUVIAL ORIGIN.

**AFTER
GLINKA,
MARBUT
AND OTHERS.**

(gleization), (4) calcification, and (5) salinization, solonization, and solodization. From the standpoint of the forester, soils developed as a result of the first three processes are of primary interest. Much of the information in the sections that follow has been obtained from Byers *et al.* (1938), Baldwin *et al.* (1938), and various soil survey reports.

PODZOLIZATION

The term podzolization is generally applied to the process or processes which lead to depletion of bases, development of acidity, and formation of eluvial *A* horizons and illuvial *B* horizons in soils. Colloids and iron and aluminum compounds are removed from the *A* horizons and accumulated in the *B* horizons. Regions where podzolization is the dominant soil-forming process are characterized by a relatively cool, humid climate and forest vegetation. The process operates with varying intensity in temperate and tropical climates wherever rainfall is sufficient to cause percolation of water through soils having acid organic matter. It is active in soils supporting hardwoods, as well as in those supporting conifers.

The most striking results of podzolization are seen in mature podzol soil profiles, for there the process has operated with greatest intensity. For this reason the chemical and physical properties of podzol soils will receive particular attention in this discussion. It should be clearly understood, however, that podzolization does not always lead to development of podzols. In the discussion that follows it will become evident that several other zonal soils, together with certain intrazonal soils, also owe their characteristics to the influence of this process.

Podzolization results in the development of three zonal soils which are relatively light-colored and under natural conditions support forest: (1) podzols, (2) brown podzolic soils, and (3) gray-brown podzolic soils. The broad features of the geographical distribution of these soils are indicated in Figures 38 and 39.

In the United States *podzol soils* occupy large areas in northern New England and the northern Lake States. They occur along the middle and north Atlantic seacoast as well as in the mountainous regions of the West. Extensive areas of podzol soils are also known in Canada and Alaska.

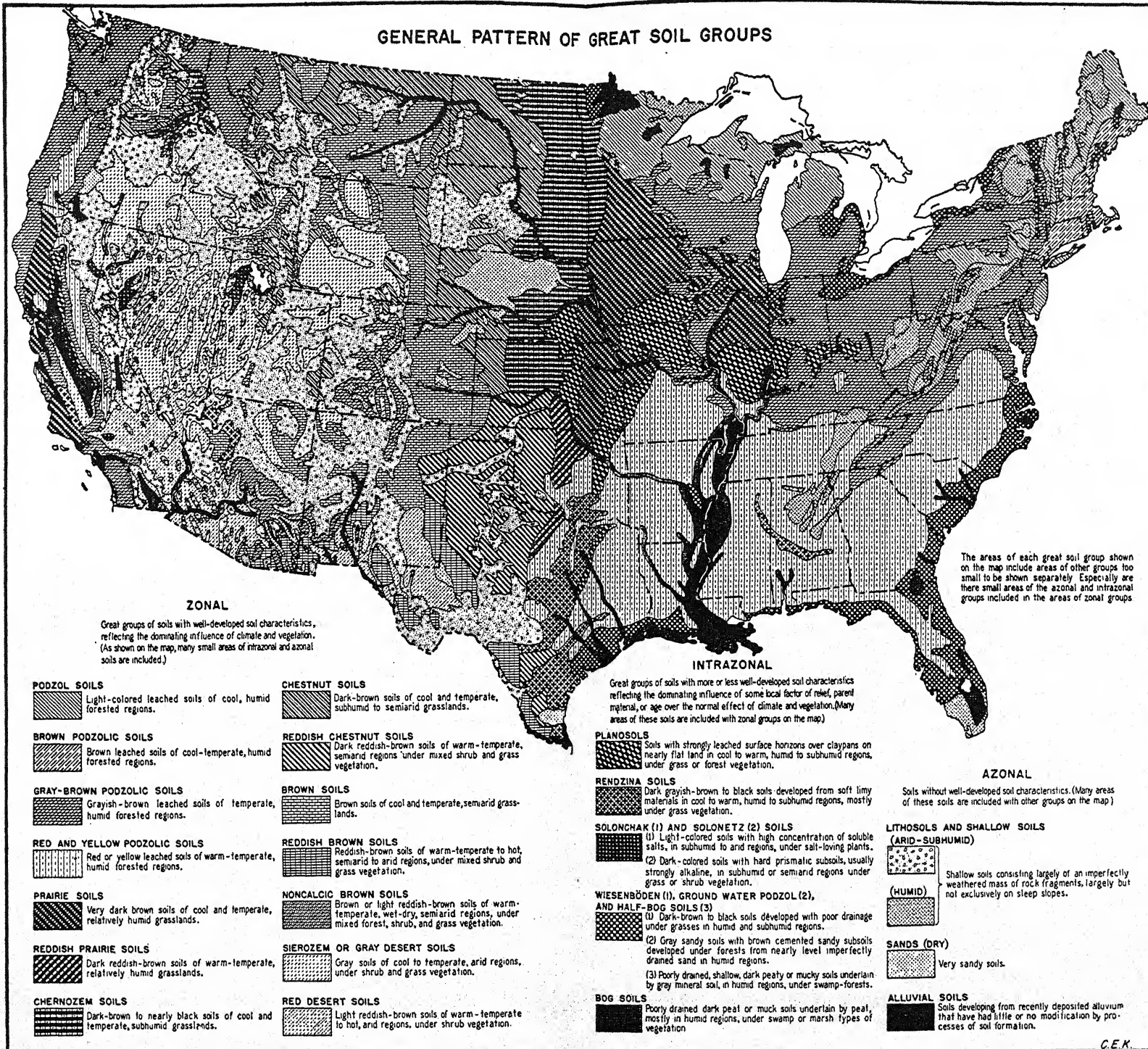
Under forest conditions podzol profiles characteristically have

a layer of acid organic matter on the mineral soil surface. The tree leaves and other organic debris reaching the soil are relatively poor in bases; and on decomposition, principally by fungi, they give rise to acid products. As the acids generated in the organic layers are moved downward by percolating water, they attack the soil minerals, leach out exchangeable cations, and generally decrease the pH of the mineral soil material. The acid weathering of the soil minerals causes a marked loss of metallic cations, including iron and aluminum, but most of the silica remains behind. Organic and inorganic colloids are dispersed and moved downward in the percolating waters. The hydrogen-saturated organic colloids appear to exert a protective influence on the inorganic colloids, keeping them from flocculating and forming aggregates. Unlike the other common soil-forming minerals, quartz is particularly resistant to attack by acid solutions. Consequently, its relative proportion in the surface layer of mineral soil increases as leaching progresses. With the removal of iron compounds and other coloring materials and with an increase in the relative proportion of quartz in the surface layer of mineral soil, where leaching is most intense, a gray color develops. This gray layer constitutes the A_2 horizon of the podzol soil profile and represents the zone of maximum leaching.

Immediately below the A_2 horizon iron and aluminum in varying degrees of oxidation and hydration accumulate, along with colloidal organic matter. According to the hypothesis of Mattson and Gustafsson (1934, 1937), these three materials form complexes with each other; they have amphoteric properties and possess different isoelectric points. Colloidal organic matter is precipitated in the upper part of the B horizon because its isoelectric point is lowest. Somewhat lower in this same horizon the zone of greatest iron accumulation is encountered, and still lower is the zone of greatest aluminum accumulation, their positions corresponding with increasingly higher isoelectric points. Although it is not a certainty that these views of Mattson and Gustafsson account fully for the accumulation of organic colloids and iron and aluminum compounds in the B horizon of podzol soils, it can be stated that chemical analyses indicate the differential accumulation already described.

Other investigators have suggested different explanations for the deposition of materials in the B horizon of podzol soils. It

GENERAL PATTERN OF GREAT SOIL GROUPS



has been thought that deposition results partly from the flocculation of the colloids by bases; in practice this hypothesis seems to differ little from the suggestion of Mattson and Gustafsson (1934). A further explanation of deposition that has been advanced is that organic substances responsible for reduction of iron in the *A* horizon become oxidized as they move downward. The iron then changes from the mobile ferrous state to the relatively immobile ferric condition. It is also believed by some workers that deposition in the *B* horizon is in part a purely mechanical process. As the porosity of the horizon is reduced, it serves as a filter to remove suspended materials from the percolating waters. It is not impossible that all these suggested mechanisms may contribute, in varying degree, to the formation of *B* horizons in podzol soils.

Aaltonen (1939, 1940) has observed that the inorganic colloids are precipitated at a deeper level in young forest soils than in more nearly mature soils. Both he and Stebutt (1930: 136) maintain that as the podzol soil profile develops, the *B* horizon "grows" from the bottom up. They are supported in this view by Mattson and Gustafsson (1937).

In the *C* horizon, below the *B*, there is no noticeable deposition of material. Since percolating waters pass through this horizon, some slight deposition and removal probably occur, but the rate of change is so slow as to be negligible.

In the preceding outline of the podzolization process as it concerns the development of podzols, no mention was made of the criteria that may be used to distinguish strongly podzolized from weakly podzolized soils. It seems evident from examination of various analytical data that the silica-alumina ratios of the colloids separated from the *A* and *B* horizons can serve as a useful criterion for this purpose. Thus in well-developed podzols (strongly podzolized soils) the silica-alumina ratio of the colloids in the *A* horizon is approximately three or more, whereas the ratio in the *B* horizon is usually less than one. There is no intent to imply that the silica-alumina ratio must always be determined before it is possible to identify a soil as a podzol. Well-developed profiles, such as those illustrated in Figures 40 and 41, are readily recognized as podzols in the field without chemical determinations of any kind. The characteristic features seen in the field are the layer of unincorporated organic matter, the gray

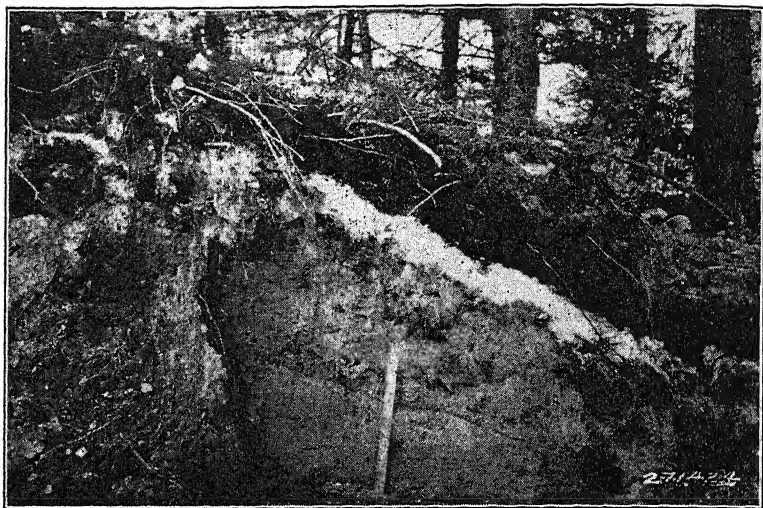


FIG. 40. A podzol soil profile, Breadloaf Township, Vermont. The white A_2 horizon is 4-5 in. thick. Photograph by United States Forest Service.

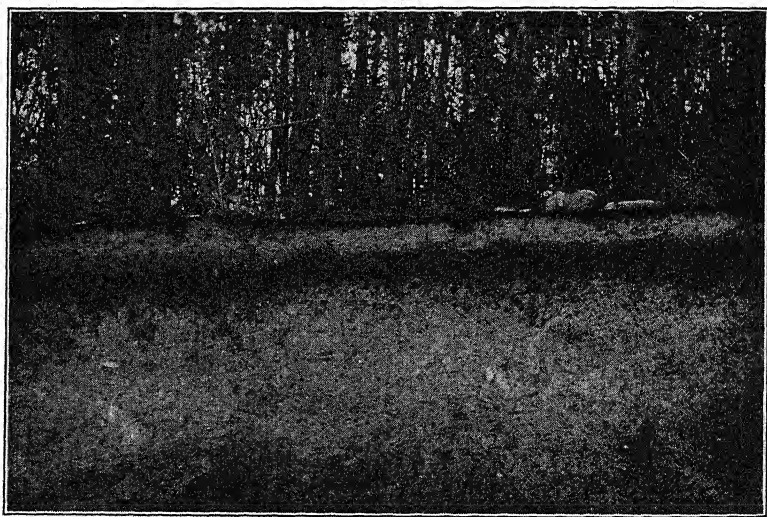


FIG. 41. A podzol soil profile developed from Keewatin calcareous till, northern Minnesota. The surface soil is a clay loam. The A_2 horizon is about 10 in. thick, the B horizon about 12 in. The present stand is aspen. Photograph by C. C. Nikiforoff.

A_2 horizon, and the brown to dark-brown B horizon. In less clear-cut profiles, however, the identity of the soil may not be obvious, and then the silica-alumina ratio of the colloids becomes critical.

A generalized description of the profile of a podzol soil may be helpful at this point. As was mentioned earlier, the first layers encountered usually consist of organic matter in varying stages of decomposition. Fairly well-defined subdivisions of the unincorporated organic matter are frequently distinguishable; they are referred to as the A_{00} and A_0 layers or as L , F , and H layers. The thickness of the unincorporated material varies from less than 1 to more than 12 in. It ranges in color from brown to black, possesses a relatively high cation-exchange capacity, and is acid in reaction. The mineral soil immediately below the humus layer is also acid in reaction and is commonly dark gray because of organic material which has washed in. When present, this soil is designated as the A_1 horizon. Below the A_1 is the A_2 horizon, in which eluviation has been most active; sometimes the A_1 horizon is lacking or is present only as a trace. The A_2 is gray or whitish gray, structureless or sometimes laminated, acid in reaction, high in silica, and relatively low in iron, aluminum, and bases. Usually it contains less nitrogen and organic matter and has lower cation-exchange capacity than the horizon below. The thickness of the A_2 varies from less than 1 to more than 30 in., depending on the degree of profile development. Underlying the bleached A_2 horizon, sometimes called *Bleicherde*, is the B_2 horizon or zone of illuviation. The B_1 horizon, which is transitional, is commonly lacking in podzols. As a rule, the B_2 horizon, which varies from about 1 to 10 in. in thickness, is brown or dark brown in color and may or may not be indurated. When firmly cemented, this material is referred to as hardpan or *Ortstein*; if indurated but not firmly cemented, it is called *Orterde*. Variations in the amount of humus and iron have resulted in recognition of so-called humus podzols and iron podzols. In passing from the B_2 to lower horizons, the color becomes lighter. The B horizon contains a relatively high content of organic matter, iron, and aluminum; occasionally there is an enrichment in bases. In podzol soils the entire profile above the C horizon has an acid reaction, but the lowest pH values generally occur in the H layer or in the A_1 horizon.

Chemical and physical data for a well-developed podzol soil profile from the Adirondack region in New York are presented in Table 39. The soil was a Beckett sandy loam and supported a virgin stand of sugar maple, yellow birch, American beech, red spruce, and balsam fir. Although the content of clay in this soil is very low, a slight tendency toward accumulation in the *B* horizons may be noted. Organic matter and the associated moisture-equivalent and cation-exchange values are highest in the *A*₀ horizon dropping to a low level in the *A*₂ but again rising in the *B* horizons. The values for silica, iron, and alumina content are characteristic. Silica content is high and iron and alumina contents are low in the *A*₂ horizon; in the *B* horizons the reverse condition is found. Most significant is the high silica-alumina ratio of 3.41 in the colloids of the *A*₂ horizon and the low ratio of 0.81 in the *B*₂ horizon.

TABLE 39. CHEMICAL AND PHYSICAL PROPERTIES OF A WELL-DEVELOPED PODZOL SOIL PROFILE FROM THE ADIRONDACK REGION

	<i>A</i> ₀ (<i>H</i> layer)	Horizon		<i>B</i> ₃	<i>C</i>
		<i>A</i> ₂	<i>B</i> ₂		
Sand content, percentage	..	78.0	80.0	75.5	68.0
Silt content, percentage	..	19.2	16.2	19.5	29.2
Clay content, percentage	..	2.8	3.8	5.0	2.8
Organic matter content, percentage	85.0	2.25	20.6	12.0	0.93
Moisture equivalent, percentage	132.9	23.05	55.07	27.18	12.76
Cation-exchange capacity, m.e. per 100 gms.	148.2	8.0	56.0	29.3	2.6
SiO ₂ content, percentage	..	86.70	67.09	75.68	77.11
Fe ₂ O ₃ content, percentage	..	1.36	3.44	2.59	2.72
Al ₂ O ₃ content, percentage	..	5.92	14.45	11.57	10.54
SiO ₂ content of colloids, percentage	..	41.45	11.75	16.95	31.02
Al ₂ O ₃ content of colloids, percentage	..	20.14	24.49	29.81	28.23
Silica-alumina ratio of colloid	..	3.41	.81	.96	1.86
pH	3.45	4.03	4.35	4.86	5.27

For further details concerning the genesis and morphology of podzol soils the discussions of Stremme (1930), Joffe (1931, 1933), Latimer (1933), and Aaltonen (1939) should be consulted.

Podzolization, as has been stated, is the dominant process in

the development of a number of soils other than podzols. It may be pointed out here that the general characteristics of these soils, to be considered presently, show points of similarity to podzols but, because of less intense leaching, do not exhibit as marked differences between horizons.

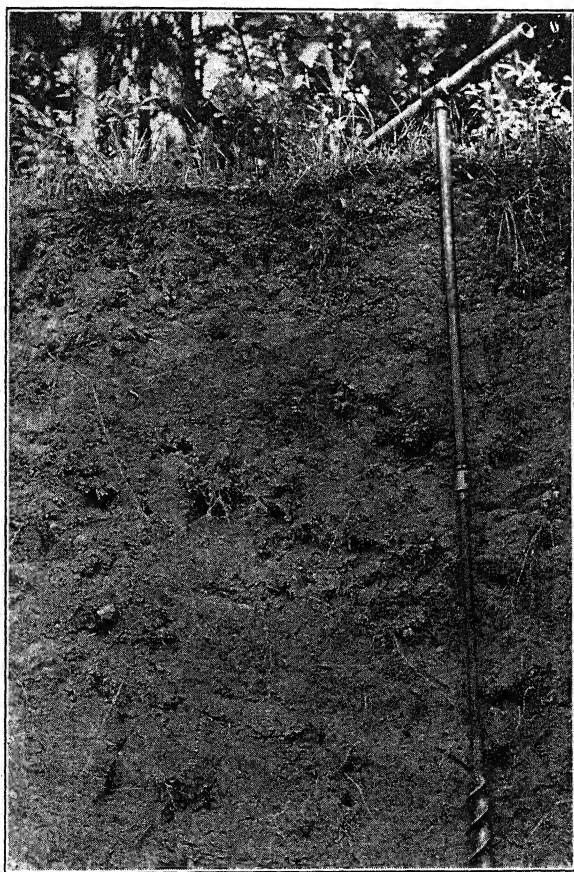


FIG. 42. A brown podzolic soil in Connecticut, showing but little horizon development. Photograph by H. A. Lunt.

The principal occurrence of *brown podzolic soils* is in southern New England, south of the region of podzols. Developed under less intense podzolization, these soils differ from podzols in having a thin, dark-gray A_1 and a thin, gray-brown or yellowish-brown A_2 overlying a brown or dark-brown B horizon (Figure 42).

The *B* is only slightly heavier in texture than the *A*. As a general rule, although not invariably, these soils support deciduous or mixed deciduous-coniferous forests. The solum (*A* + *B* horizons) is seldom more than 24 in. thick. Data presented by Brown and Byers (1938) indicate that the chemical composition of the soil material in the various horizons of brown podzolic soils is considerably more uniform than in podzols. Some leaching of bases, particularly calcium, occurs, but translocation of colloids is usually slight. The content of organic matter and nitrogen is greatest in the *A* horizon and decreases sharply with increasing depth in the soil body. The silica-alumina ratio of the A_1 horizon colloid of Gloucester sandy loam from Massachusetts is 2.48, and of the B_2 horizon colloid 0.79. Comparable ratios representing Gloucester sandy loam obtained from New Jersey are 1.72 and 0.85. Although podzolization is less severe in these soils than in podzols, it is evident that there has been a substantial fractionation of the colloids during profile development. For further morphological and chemical descriptions of brown podzolic soils the work of Lunt (1932) should be consulted.

Gray-brown podzolic soils occur chiefly in regions south of the podzols and the brown podzolic soils. They occupy large areas in the eastern and midwestern parts of the United States and occur in northwestern Montana, northern Idaho, and on the west side of the Cascades in Washington and Oregon. The genesis and morphology of these soils have been discussed in some detail by Baldwin (1928) and Veatch (1928). Although developed as a result of the podzolization process, profiles of gray-brown podzolic soils show less clearly defined horizons than podzol soils. The less intense leaching of the gray-brown podzolic soils must be attributed to the more temperate climatic conditions under which they were developed.

Profiles of gray-brown podzolic soils generally exhibit a relatively thin layer of unincorporated organic matter. The A_1 horizon is grayish brown because of incorporated organic matter; this layer is generally 2-4 in. in thickness. The A_2 horizon is grayish brown or light yellow, and the somewhat heavier-textured *B* horizon is brown, yellowish brown, or reddish brown. There is an apparent accumulation of iron and aluminum, but not of organic matter, in the *B* horizon (Marbut, 1935: 22). The solum seldom exceeds 4 ft. in thickness. Byers, Alexander, and Holmes

(1935) have presented chemical data for Miami silt loam which may be regarded as typical of soils of the gray-brown podzolic group. The content of silica is higher in the *A* than in the *B*; on the other hand, iron and alumina have been removed from the *A* and accumulated in the *B*. Another characteristic feature is the loss of calcium from the upper soil horizons by leaching. The distribution of nitrogen in gray-brown podzolic soils is similar to that in brown podzolic soils. It is noteworthy that a pronounced textural difference exists between the *A* and the *B* horizons. The clay content (< 0.002 mm.) in the A_2 horizon of Miami silt loam was reported by Byers, Alexander, and Holmes as 20.2 per cent, that in the *B* horizon as 50.6 per cent. A marked downward movement of clay during profile development is indicated. The silica-alumina ratio of the colloid remains essentially the same throughout the profile, the values being 3.48 in the *A* horizon and 3.41 in the *B*. It may be concluded that one of the important chemical differences between gray-brown podzolic soils and podzols is that in podzols there is an extreme fractionation of the colloids, whereas in the gray-brown podzolic soils this tendency is much less marked, the clay accumulation in the *B* horizon having resulted from downward migration of the whole colloid.

LATERIZATION

The process of laterization is most active in forested regions characterized by humid tropical climates. Under these conditions chemical weathering is very intense; soil minerals and the parent rocks are subjected to extremely strong hydrolysis and oxidation, with the release of bases and the production of silicic acid and hydrated oxides of iron and aluminum. In the initial stages of the genesis of laterites there is a rapid liberation of bases such as calcium, magnesium, potassium, and sodium, resulting in nearly neutral or even alkaline conditions in the upper layers. This initial condition of low acidity, or even alkalinity, is held to be further favored by the absorption and return to the surface by plants of at least a part of the liberated bases. Rapid decomposition of the organic matter results in the release of the contained bases. With the reaction nearly neutral or alkaline, solution and removal of silica from the silicate minerals is favored; iron, aluminum, and manganese oxidize and are left behind, together with the quartz-silica. Fox (1936) has in-

licated that even in quartz, a mineral which is exceedingly resistant to weathering, some desilicification occurs under laterization. The base content is progressively reduced by leaching with the result that in time, after hydrolysis of the primary minerals is complete, an acid condition develops; podzolization may then occur.

Laterization is regarded by some soil scientists more as a geological process of rock weathering than as a process of soil development (Kellogg, 1936; Byers *et al.*, 1938; Bonnet, 1939; and others). The view is held that laterization results in the formation of parent material which, as a result of podzolization, gives rise to red podzolic and yellow podzolic soils.

Pendleton (1936, 1940) has urged that the term *laterite* be confined to the original meaning given by Buchanan in his Journal, written in 1800-1801. Excerpts from Buchanan's work are available in a report by Fox (1936), who studied laterite in Malabar, India, the type area where the term was first used. It seems clear that originally the term *laterite* was applied to a highly ferruginous, usually vesicular, clayey material which hardens to the consistency of rock on exposure to air. In fresh condition the material is relatively soft and easily cut into blocks, which after hardening form a serviceable building material widely used in the tropics. Laterite is viewed by Pendleton as an illuvial sesquioxide horizon which forms in well-developed soils of the humid tropics; it is at the surface only if erosion has truncated the profile.

Actually the term *laterite* is used by many, if not most, American soil scientists to denote a "zonal group of soils having very thin organic and organic-mineral layers over reddish leached soil that rests upon highly weathered material, relatively rich in hydrous alumina or iron oxide, or both, and poor in silica; usually deep red in color."¹ This concept of a laterite, although obviously broader than that favored by Pendleton, is employed by the present authors because of its common usage in the general classification of tropical soils currently standard in America. For the same reason the term *lateritic*, employed to designate soil or soil material having the characteristics of a laterite, has been retained.

¹ From a glossary of special terms used in *Soils and Men*, Yearbook of Agriculture, 1938. United States Department of Agriculture. Pp. 1162-80.

The laterization process, in conjunction with podzolization, has resulted in the development of six zonal soils in the forested warm-temperate and tropical regions: (1) laterites, (2) reddish-brown lateritic soils, (3) yellowish-brown lateritic soils, (4) red podzolic soils, (5) yellow podzolic soils, and (6) terra rossa. Ground-water laterites are intrazonal soils. The general features of the geographical distribution of these soils are indicated in Figures 38 and 39.

Laterites are not known in the United States, but they occur in Puerto Rico, the Hawaiian Islands, and other tropical regions. As a result of extremely rapid decomposition there is little tendency for the development of layers of unincorporated organic matter on the mineral soil surface. The mineral soil horizon containing incorporated organic matter is likewise thin. Below this horizon is a reddish, leached soil which rests on very highly weathered material. This material is deep red or red and may be reticulately mottled with red and light gray, buff, or white. It is relatively rich in iron and aluminum and poor in silica. As a result of the very intense weathering, laterite soils reach a depth of 100 ft. or more. Laterite soils are characterized by good to excessive drainage, low plasticity, low erosivity, and low fertility. Much, if not most, of the mineral nutrient capital of a site in the tropics having laterite soil is represented by the vegetation itself. Consequently, removal of the vegetation results in loss of a very substantial proportion of the total supply of mineral nutrients. Laterites, because of their low fertility, are sometimes referred to as dead soils. They have a low cation-exchange capacity and are exceedingly poor in bases.

Outstanding chemical characteristics of laterite soils are the low content of silica and the high content of iron, aluminum, and manganese. The silica-alumina ratio of colloids from both the A and B horizons is commonly about 1.00. In their investigations in Sierra Leone, Martin and Doyne (1927) separated laterite soils and lateritic soils on the basis of the silica-alumina ratio of the colloid in the first foot of soil, presumably the A horizon. In laterite soils the ratio was less than 1.33, and in lateritic soils it was 1.33-2.00. For additional information on laterite and other tropical soils the writings of Harrassowitz (1930), Anon. (1932), and Thorp and Baldwin (1940) should be consulted.

Reddish-brown lateritic soils and *yellowish-brown lateritic soils* are known to occur in both Puerto Rico and the Hawaiian Islands. The reddish-brown lateritic soils have dark reddish-brown granular surface soil, red friable clay *B* horizons, and red or reticulately mottled lateritic parent material. The yellowish-brown lateritic soils are characterized by yellowish-brown, friable, and granular surface horizons high in clay content, under which is yellow or reddish-yellow friable clay material overlying parent materials not strongly mottled.

In the broad transition zone between the laterite soils of the humid tropics and the gray-brown podzolic soils of humid temperate regions are developed the lateritic red podzolic and yellow podzolic soils. The yellow and red podzolic soils differ from the gray-brown podzolic soils in other respects than color; they have a lower percentage of bases and silicate silica and a higher percentage of iron and aluminum. The yellow and red soils are more thoroughly eluviated, and there is a greater difference in texture between the *A* and *B* horizons. The yellow soils occur predominantly on smooth or flat areas; the red soils are more common on rolling areas with good drainage. Both these soils owe their characteristics to both laterization and podzolization. Variations in reaction from extremely acid to only slightly acid are known.

Red podzolic soils occupy extensive areas in the southeastern United States and also occur in the West Indies. The natural vegetation consists of deciduous, coniferous, or mixed forest. These soils have a thin layer of unincorporated organic matter resting on a gray or dark brownish-gray humified mineral soil horizon (A_1), having a thickness of as much as 2-3 in. The A_2 horizon is yellowish gray or light pinkish gray, more or less sandy, and several inches thick. Underlying the A_2 is a red or brownish-red *B* horizon consisting of clay having a granular structure. This horizon, which may be 1-3 ft. thick, rests on reticulately mottled parent material.

Yellow podzolic soils seem to have developed under conditions of greater soil moisture than are found in areas of red podzolic soils. The view is frequently expressed that a previous condition of imperfect drainage may be responsible for the higher degree of hydration of iron compounds and the consequent color in yellow podzolic soils. These soils occupy large areas in the coastal

plain of the southeastern United States. A thin layer of organic matter commonly overlies the A_1 horizon, which is about 3 in. thick and dark-colored, because of incorporated organic material. The A_2 horizon is light yellow and has a thickness of one to several feet. Both the A_1 and A_2 horizons are commonly structureless. Underlying the A_2 is a yellow or light reddish-yellow heavy-textured B horizon, extending downward 3 ft or more. The parent material of the yellow podzolic soils is lateritic but generally lighter-colored than that of the red soils.

Terra rossa (red earth) is a term commonly used to designate red soils which have developed in regions having a Mediterranean type of climate; some soil scientists prefer to limit the name to red soils which have developed from limestones. The soils are usually well drained and friable and have an alkaline reaction. Native vegetation generally consists of an open stand of trees with shrubs and grasses. No examples of *terra rossa* are known in the United States.

Ground-water laterites occur in hot, humid regions in situations having poor drainage. Native vegetation consists of tropical forest. These soils are characterized by a light-colored, leached A horizon overlying a thick, reticulately mottled, and cemented hardpan. The soils are acid and contain concretions with a high content of iron, aluminum, and sometimes manganese. Fluctuations in the water-table result from the alternation of wet and dry seasons. No examples of ground-water laterites are known in the United States.

COMPARISON OF THE CHEMICAL PROPERTIES OF PODZOLIZED AND LATERIZED SOILS

The foregoing discussion of great soil groups developed under the influence of podzolization and laterization has been primarily concerned with general profile characteristics. In order to clarify and summarize the principal differences between soils developed under the influence of these two processes, certain comparisons of chemical characteristics will be presented.

Data illustrating the partial chemical composition of a podzol soil (Brassua sandy loam) from New Hampshire and of a red podzolic soil (Cecil sandy clay loam) from North Carolina are given in Table 40. These data, from the work of Brown and Byers (1938), demonstrate the high content of silica and the

relatively low content of iron and alumina in the A horizon of soils of both of these great groups. It may also be noted that there is a considerable accumulation of iron and alumina in the

TABLE 40. PARTIAL CHEMICAL ANALYSIS OF A PODZOL AND RED PODZOLIC SOIL

(After Brown and Byers, 1938)

Horizon	SiO ₂ , percentage	Fe ₂ O ₃ , percentage	Al ₂ O ₃ , percentage
<i>Podzol Soil</i>			
A ₂	86.70	1.36	5.92
B ₂	67.09	3.44	14.45
C	77.11	2.72	10.54
<i>Red Podzolic Soil</i>			
A	75.06	4.16	10.80
B ₂	57.32	11.20	22.16
C	54.50	11.27	23.05

B₂ horizon of each. However, the chemical distinctions between these two representative soils are not clearly demonstrated by the analyses in Table 40. The fundamental differences are more clearly indicated by the chemical data for the colloidal fraction supplied in Table 41; these data were also drawn from the report by Brown and Byers (1938).

TABLE 41. PARTIAL CHEMICAL ANALYSIS OF THE COLLOIDS FROM A PODZOL AND RED PODZOLIC SOIL

(After Brown and Byers, 1938)

Horizon	SiO ₂ , percentage	Fe ₂ O ₃ , percentage	Al ₂ O ₃ , percentage
<i>Podzol Soil</i>			
A ₂	41.45	4.45	20.14
B ₂	11.75	7.30	24.49
C	31.02	9.59	28.23
<i>Red Podzolic Soil</i>			
A	32.81	12.57	33.68
B ₂	35.75	16.17	32.22
C	35.13	17.54	30.52

From inspection of the data in Table 41 it is evident that there are pronounced chemical differences between the colloids in the

A and B horizons of the podzol soil, but the differences are relatively slight in the red podzolic soil. Fractionation of colloids in the podzol soil has been extreme, but in the red podzolic soil only slight. The comparatively uniform composition of colloids from the red podzolic soil is regarded as evidence that the whole colloid was moved downward during profile development. The differences in composition of material in the horizons of the red podzolic soil shown in Table 40 are evidently the result not of differences in composition of the colloid (see Table 41) but of differences in the amount of colloid. In support of this view it may be stated that the amount of colloid in the A horizon of this soil is 17.3 per cent, and in the B₂ horizon 48.6 per cent. No such increase is noted in the amount of colloidal material in the B₂ horizon of the podzol soil. Both fractionation of colloids and mechanical movement of the whole colloid occur in podzolized soils, but fractionation is regarded by the authors as the more characteristic feature in soils developed under intense podzolization. Very little fractionation of colloids is noted in laterized soils; that is, the composition of the colloid remains relatively uniform throughout the profile. An exception occurs in true laterites, where extensive removal of silica may be noted in the A horizon.

TABLE 42. SILICA-ALUMINA RATIOS OF COLLOIDAL CLAY FROM THE A AND B HORIZONS OF REPRESENTATIVE SOILS OF CERTAIN GREAT GROUPS

(After Byers *et al.*, 1935, and Brown and Byers, 1938)

Great Soil Group	Silica-Alumina Ratio	
	A Horizon	B Horizon
Podzol	3.41	0.81
Brown podzolic	2.48	0.79
Gray-brown podzolic	2.33	2.42
Red podzolic	1.65	1.88
Laterite	1.01	1.78

The silica-alumina ratio of colloids, as was mentioned earlier, is useful in characterizing soils of the great groups. In order to facilitate comparisons, data believed to be representative have been assembled in Table 42. The ratios were obtained from Byers, Alexander, and Holmes (1935) and Brown and Byers (1938).

It appears that in going from cool, humid climates, where podzolization is the dominant process, to warm, humid climates, where laterization is the dominant process, there is a gradual reduction in the silica-alumina ratio of the colloids in the *A* horizon. In the laterites there is a notable removal of silica from the *A* horizon. The ratios indicate a relative accumulation of aluminum in the *B* horizon of podzol and brown podzolic soils but not in the other soils.

FORMATION OF PEAT AND POORLY DRAINED SOILS (GLEIZATION)

Peat is formed when decomposition of plant remains is inhibited by the presence of excessive moisture. Deposits are most common in regions having cold, or cool, humid climates, but they also occur in humid, warm-temperate and tropical regions. The nature of peat depends on the type of vegetation from which it was derived and the environmental conditions under which it was accumulated. For a detailed treatment of this subject the work of Waksman (1942) should be consulted. Three major divisions of peat may be recognized: (1) the more or less gelatinous sedimentary precipitate, (2) fibrous peat, and (3) woody peat (Dachnowski, 1926). Muck is organic soil material, usually having a relatively high mineral content, which results from decomposition of peat.

The accumulation of peat and muck, together with gleization, leads to development of two intrazonal soils, namely, *bog soils* and *half-bog* or *marsh border soils*. Bog soils have a peat or muck surface soil underlain by peat; the muck or peat surface soil of half-bog soils is underlain by grayish and mottled mineral material. Both soils may support forest vegetation. Bog and half-bog soils occur in the regions of podzol, brown podzolic, gray-brown podzolic, and yellow podzolic soils.

Gleization is a process of soil formation resulting in the development of a glei (or gley) horizon in the lower part of the soil profile above the parent material. A glei horizon develops under conditions of poor drainage as a result of lack of oxygen. Reduction of the iron compounds leads to a bluish-gray, greenish, or gray color in the more or less sticky, compact, and often structureless glei-horizon material. In situations where the water-table fluctuates, oxidizing conditions may prevail during a part of the year, resulting in mottling of yellow, brown, or red

adjacent to old root canals and other openings where access to oxygen is afforded. At or near the water-table ferrous iron in solution is oxidized to the ferric state and deposited, causing formation of iron hardpan or *Ortstein*.

Ground-water podzols, various planosols, *Wiesenböden* (meadow soils), alpine meadow soils, and tundra soils, in addition to bog and half-bog soils, are developed under the influence of gleization. Tundra soils are regarded as zonal, all others as intrazonal.

Ground-water podzol soils are the result of both podzolization and gleization. They are usually developed in cool to tropical humid regions from sandy parent material in situations of imperfect drainage. These soils have a limited occurrence in the podzol and gray-brown podzolic regions but occupy extensive areas in the coastal plain of the southeastern United States. A generally thin layer of acid organic matter overlies a horizon of light-gray leached material as much as 2-3 ft. thick. Below this is a dark-brown *B* horizon with an accumulation of iron or humus or both. As a rule the *B* horizon material is cemented (*Ortstein*) or indurated (*Orterde*). The parent material is generally grayish. Ground-water podzol soils support forest vegetation.

Planosols, developed in humid or subhumid climates, have leached surface horizons underlain by *B* horizons which are more strongly illuviated, cemented, or compacted than those in associated normal soils. They are old soils which have developed upon nearly flat surfaces under conditions of imperfect drainage. Planosols occupy large areas in the Middle West and are known to occur in Texas and California. Both forest and grass vegetation occur on planosols. The dominant soil-forming processes are podzolization and gleization.

Wiesenböden or meadow soils are developed in humid to subhumid regions under grass and sedge vegetation. They resemble the chernozem and prairie soils. The upper part of the soil body is dark brown or black because of a high content of rather well-decomposed organic matter. As a rule, it is not acid. At a depth of 1-2 ft. there is a transition to grayish mineral soil mottled with brown. *Wiesenböden* are of frequent occurrence in poorly drained situations in the regions of gray-brown podzolic soils and prairie soils. They are also known in other soil regions.

Alpine meadow soils are found at and above timber line in the Rockies and the Cascades. Drainage is usually poor, and atmospheric temperature low. Native vegetation consists of grasses, sedges, and other herbaceous plants with occasional scattered trees and shrubs. The surface soil is dark brown, because of a high content of organic matter. At a depth of 1-2 ft. there is a transition to grayish, mottled soil.

Tundra soils occur in the subarctic regions of northern Europe, Asia, northern Canada, and Alaska. The vegetation consists of lichens, mosses, herbs, and dwarf shrubs. As a result of poor drainage and low temperatures a peaty surface layer is developed. Underlying this layer is grayish mineral material mottled by brown, which grades below into a gray-blue or greenish-blue horizon. The substratum is perpetually frozen.

CALCIFICATION

The term *calcification* is applied to the process or processes of soil formation in which the surface soil is kept supplied with sufficient calcium to maintain the colloids in a high state of base saturation. Calcification is the dominant soil-forming process in temperate-zone regions having less than about 25 in. of precipitation annually and in tropical regions having less than about 45 in. The native vegetation consists principally of grasses or shrubs.

The high calcium content of soils developed as a result of calcification is explained by the combined influence of low rainfall and native vegetation. Some leaching of soluble materials from the surface soil occurs, but the amount of water involved is insufficient to cause their removal from the soil body. Consequently an accumulation of calcium and magnesium carbonates is commonly found at approximately the average depth to which percolating waters penetrate the soil profile. Maintenance of a high base content is also favored by the grass vegetation, which absorbs relatively large amounts of bases from the lower horizons and returns them to the surface soil layers.

As a result of the calcification process, soil reaction is nearly neutral; the colloids, largely saturated with calcium, are relatively immobile. This situation accounts for the fact that the movement of colloids from the *A* to the *B* horizon is slight. The microorganic population is predominantly bacterial.

Calcification is the dominant process in the development of the following zonal soils: (1) chernozem, (2) chestnut, (3) reddish chestnut, (4) brown, (5) reddish brown, (6) sierozem, (7) desert, and (8) red desert. The first three zonal soils are dark-colored and occur in subhumid to semiarid regions; the remainder are light-colored and are found in semiarid to arid

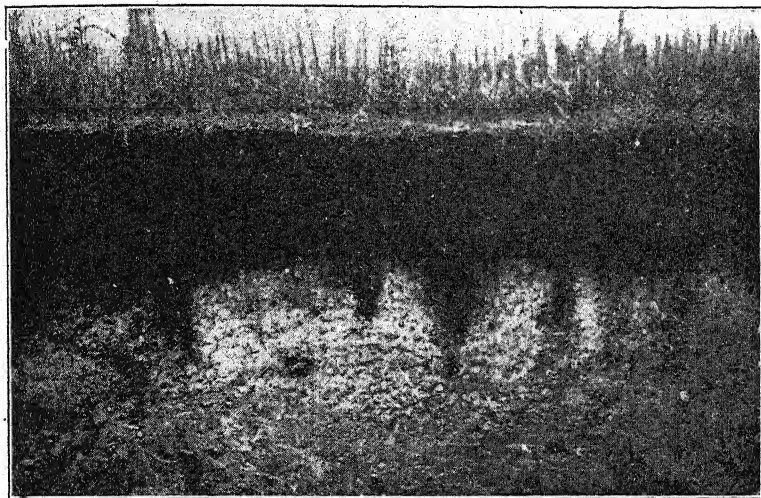


FIG. 43. A chernozem soil profile developed from Keewatin calcareous drift in Minnesota. The dark-colored A horizon is about 12 in. thick. Compare with Figure 41. Photograph by C. C. Nikiforoff.

regions. Four additional zonal groups of soils are recognized as having developed under the joint influence of calcification and podzolization in subhumid to humid regions, namely, prairie, reddish prairie, degraded chernozem, and noncalcic brown. Intrazonal soils developed as a result of calcification are represented by the brown forest soils and rendzina soil groups. The general distribution of these soils is indicated in Figures 38 and 39.

Chernozem (also spelled *tschernosem* or *tschernosiom*) soils are developed in temperate to cool, subhumid regions. They occur in the most humid parts of regions having soils with a layer of calcium carbonate accumulation. The principal occurrence of chernozem in the United States is in a broad belt lying immediately west of the prairie soils. Tall and mixed grasses constitute the native vegetation. A chernozem profile is charac-

terized by a nearly black, friable surface soil, high in organic matter and extending to a depth of as much as 3–4 ft. (Figure 43). It has a slightly acid or slightly alkaline reaction. Below the dark-colored surface soil is a lighter-colored horizon transitional to the horizon of calcium carbonate accumulation.

Chestnut soils occur in temperate to cool, semiarid regions, such as the central and northern part of the Great Plains. They were developed under conditions of lower rainfall than chernozems and support a native vegetation of mixed tall and short grasses. The profile usually consists of a dark-brown, platy surface horizon which grades below into a layer of brown, prismatic soil and finally, at a depth of 1.5–4.5 ft., into a horizon of calcium carbonate accumulation.

The *reddish chestnut soils* were developed in warm-temperate to hot, semiarid regions having a native vegetation of mixed grasses and shrubs. In the United States the principal occurrence of soils of this group is south of the area of chestnut soils and west of the reddish prairie soils. The reddish-brown surface soil extends to a depth of as much as 2 ft. It grades below into reddish-brown or red soil of heavier texture and finally into the horizon of lime accumulation.

Brown soils occur in temperate to cool, arid to semiarid regions. The native vegetation consists of short grasses and bunch grasses. In the United States brown soils form a wide belt lying to the west of the chestnut and reddish chestnut soils. The brown surface horizon grades into lighter-colored soil and finally into a whitish horizon of calcium carbonate accumulation at a depth of 1–3 ft.

Reddish-brown soils are developed in arid to semi-arid climates in warm-temperate to tropical regions. The native vegetation consists of a rather thin cover of short grasses and shrubs. In the United States reddish-brown soils occur from western Texas to southern Arizona. The surface horizon is light brown with a slightly reddish cast, grading into a dull reddish-brown or red heavier-textured subsoil. Below this subsoil, at a depth ranging from 2–6 ft., is a whitish layer of lime accumulation, which may or may not be cemented.

Sierozem or *gray desert soils* are developed in semidesert regions characterized by temperate to cool, arid climates. They cover large areas in the intermountain sections of the western

United States. The native vegetation consists principally of desert shrubs and grasses. The surface horizon is brownish gray, grading through lighter-colored soil into calcareous material,

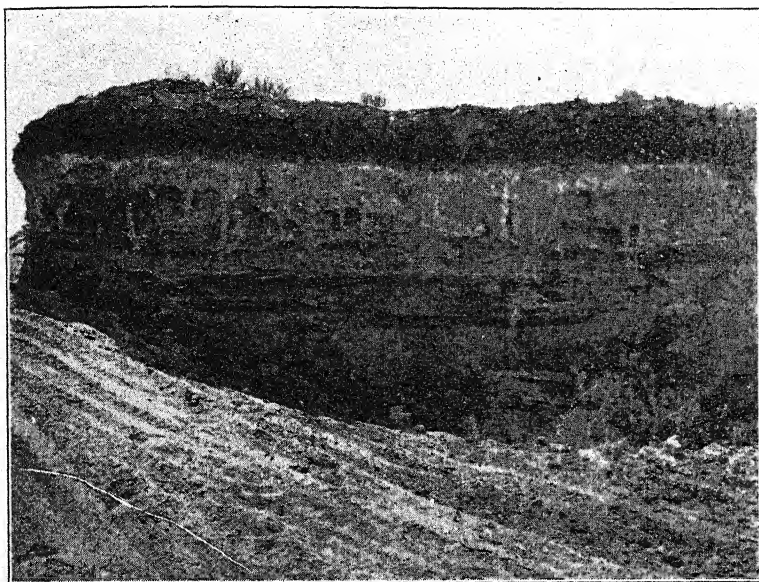


FIG. 44. A sierozem soil profile in southern Idaho, showing calcium carbonate accumulations.

which represents the layer of carbonate accumulation. The horizon of lime accumulation, occurring at a depth of about 1 ft. or less, frequently forms a cemented hardpan (Figure 44).

Desert soils occur in temperate to cool, arid climates; they occupy vast areas in the western United States. Native vegetation, which is very scanty, consists principally of desert shrubs. The surface soil is light gray or light brownish gray and is underlain by calcareous material, which is frequently cemented to hardpan. This hardpan is commonly called *caliche*. Wind erosion is severe in desert areas because of the scanty cover of vegetation. The fine soil particles are removed from the surface, and the coarser material, which cannot be moved by the wind, remains to form a desert pavement.

Red desert soils are developed in warm-temperate and tropical regions having an arid climate. In the United States they are

known to occur in the Southwest from Texas to southeastern California. Desert shrubs form the principal native vegetation. The surface soil is light reddish brown and is underlain with a reddish-brown or red, heavier-textured subsoil. This subsoil grades into a horizon of lime accumulation, which may or may not be cemented.

Prairie soils are developed as a result of calcification with weak podzolization. They occur in regions having a cool, moderately humid climate, similar to that which characterizes the regions having gray-brown podzolic soils. The native vegetation, however, consists of tall grasses, and not forest. Typical prairie soils occupy extensive areas in the Middle West, east of the region of chernozem soils and west of the forested podzol, gray-brown podzolic, and red and yellow podzolic soils; prairie soils also occur in California and in the Northwest.

The prairie soils, developed under relatively humid climatic conditions, have no horizon of calcium carbonate accumulation. In this respect they differ from the chernozems which are developed in drier regions. Although some leaching of prairie soils occurs, the native grasses return sufficient calcium to the surface to prevent appreciable unsaturation of bases and dispersion of colloids. The surface horizon is very dark brown or grayish brown and is well supplied with organic matter and bases. Below it the soil grades through brown to the lighter-colored parent material at the depth of 2-5 ft.

Reddish prairie soils, like the true prairie soils, are developed as a result of calcification and weak podzolization. These soils occur in regions having a warm-temperate, humid to subhumid climate and native vegetation consisting of tall grasses. In the United States the reddish prairie soils lie south of the zone of the true prairie soils. The reddish color is in part inherited from the parent material and in part due to the relatively high temperatures under which these soils have developed. Surface horizons are dark brown or reddish brown, grading into reddish heavier-textured subsoil. The soils are slightly to moderately acid.

Degraded chernozem soils occupy relatively small areas in the zone between chernozems and podzols in northwestern Minnesota; they are also known to occur above the chernozems on mountain slopes. Their occurrence in Russia has been discussed

repeatedly. These soils develop when chernozem which supports grasses is invaded by forest (Glinka, 1924; Tiurin, 1930). Under grass vegetation calcification is the dominant soil-forming process, but with the appearance of trees podzolization occurs. As the vegetation changes from grass to forest, the incorporated organic matter of the soil decreases, and the upper horizons become lighter in color and more acid. The A_1 horizon is nearly black, the leached A_2 horizon grayish, and the B horizon dark brown and usually heavier textured. The A horizon is acid, but remnants of a layer of lime accumulation are found in the deeper horizons.

Noncalcic brown soils are the result of calcification with weak podzolization. They are developed in southern and central California and central Arizona in sections having a warm, semi-arid or subhumid climate with cool, moist winters and hot, dry summers. Natural vegetation usually consists of chaparral or an open stand of deciduous and coniferous trees [digger pine (*Pinus sabiniana*) and ponderosa pine] with grasses and shrubs. The A horizon is brown or light brown and slightly acid; the B horizon is reddish brown or dull red and heavier textured.

Brown forest soils (*Braunerde* of the Germans) are the result of calcification with very little podzolization. They are intrazonal soils which occur in humid, cool-temperate to warm-temperate regions, having developed as a rule under deciduous forest from parent materials rich in calcium. Brown forest soils are of very local occurrence in the United States in the region of gray-brown podzolic soils but occupy considerable areas in north central and western Europe (Glømmé, 1928; Stebutt, 1929). They usually support hardwoods which absorb large amounts of calcium, returning most of it to the surface soil in the leaf fall. The surface soil is maintained in a nearly neutral condition and is very fertile. It is dark-brown and granular and contains much incorporated organic matter. The B horizon is lighter colored and shows little or no illuviation of iron and aluminum; at a depth of 2-4 ft. the C horizon is encountered. Brown forest soils in Europe are reported to be readily influenced by forest management practices (Lundblad, 1924; Burger, 1926; Ganssen, 1937).

Rendzina soils are intrazonal and are developed under the influence of calcification. They occur in a great variety of

climatic regions, varying from cool to hot and from humid to semiarid. The factor chiefly responsible for their development is parent material, such as chalk, marl, and soft limestone, having a very high content of readily available calcium. Even in humid regions where climatic conditions favor forests, the native vegetation of rendzina soils is principally grass. This is well illustrated in the so-called Black Belt of Alabama and Mississippi, where the Sumter, Houston, and Bell soils, derived from the soft Selma chalk, originally supported grassland vegetation. Other areas of rendzina soils are known in east central and southern Texas and in the coastal region of California. The granular surface soil is usually brown or black and is underlain by grayish or yellowish calcareous material. As a rule, there is a sharp distinction between the dark-colored soil and the light-colored parent material. In reaction rendzina soils are generally either nearly neutral or alkaline. In humid climates rendzina soils may undergo leaching with ultimate transformation to zonal soil types, such as podzols.

SALINIZATION, SOLONIZATION, AND SOLODIZATION

The intrazonal solonchak, solonetz, and soloth soils are developed under the influence of salinization, solonization, and solodization, respectively. They occur in the subhumid to arid regions of the western United States and usually support a sparse cover of plants, ranging from halophytes to xeric shrubs and grasses; forests are lacking. These soils are commonly referred to in America as alkali soils.

Solonchak soils, also referred to as salty or saline soils, develop as a result of the accumulation of large amounts of soluble salts of sodium, calcium, magnesium, and potassium. The salts occur in the form of chlorides, sulfates, carbonates, and bicarbonates in variable proportions in different soils. Sodium chloride is frequently the most abundant salt. The term *white alkali* is a popular designation for the chlorides and sulfates of sodium, calcium, and magnesium. Conditions for salinization or the accumulation of soluble salts are most frequently afforded in arid or semiarid regions at places where drainage is imperfect. Solonchak soils occupy large areas on the old lake flats and alluvial bottom lands in the Great Basin in Utah and Nevada. A gray, thin, salty crust covers the surface of solonchak soils.

Below this crust is fine granular soil, which is underlain by grayish, friable, salty soil. The granular condition of the soil is caused by the flocculating effect of the soluble salts. Reaction is slightly or strongly alkaline. The profile characteristics of specific soils may vary materially from those just described.

Solonetz or black alkali soils develop as a result of solonization, which involves desalinization and alkalization of solonchak soils. With improved drainage of a solonchak soil the excess of soluble salts is leached out, the colloids become dispersed, and a strongly alkaline reaction develops. Dispersal of the colloids and development of a high degree of alkalinity result from hydrolysis of the sodium to form sodium hydroxide. Sodium carbonate is also formed, and under its influence the organic matter is dissolved and dispersed, giving rise to a dark or black soil color. The very thin, friable surface soil is underlain by dark, heavy-textured material having a columnar structure. Solonchak soils containing large amounts of alkaline-earth salts such as calcium sulfate and calcium chloride may, on removal of the salts under improved drainage, revert in the course of time to the normal soil for the region. In this event solonization, with development of a solonetz soil, is avoided.

Soloth soils develop from solonetz soils as a result of the process referred to as solodization. This process involves eluviation of the dispersed colloids with development of an acid A horizon, having little colloidal material, over a heavy-textured B horizon. Conditions for the development of grasses are materially improved by the dealkalization.

DEVELOPMENT OF AZONAL SOILS

Azonal soils are developed in all climatic regions. Consequently no single soil-forming process can be regarded as characteristic of azonal soils as a group; each of these soils is influenced by the soil-forming process dominant in the region where it occurs. Characteristics of azonal soils are primarily due to the nature of the parent material, and not to processes such as podzolization, laterization, or calcification. Soils in this group are usually so young that the processes of development have not yet resulted in well-defined profile features.

Dry sands lacking appreciable profile development are regarded

as azonal soils. Shifting dunes and dune sands only recently stabilized by vegetation are familiar examples of these soils. They are known to occur locally at many points along the Atlantic and Pacific coasts, along the eastern shore of Lake Michigan, and in the arid and semiarid regions of the West. In humid or subhumid regions forest vegetation develops on stabilized dune sands. The Landes region in France and the Nebraska National Forest furnish interesting examples of the utilization of these soils in forestry. Excessive drainage, low field capacity, and susceptibility to wind erosion when exposed characterize dune sands.

Lithosols, or skeletal soils, are most frequently encountered on steep slopes. They occupy extensive areas in the mountainous and hilly regions of the United States. In most of the national forests of the West these soils are extremely common. A thin layer of stony soil material usually overlies bedrock; erosion is generally active, removing the products of weathering about as fast as they are produced. In arid regions lithosols may be barren or support a scanty cover of grasses and shrubs, but in humid regions they support forest vegetation. Because of the shallow soil body, windthrow is common and timber growth relatively poor. However, good stands may develop in situations where the underlying rock is broken by fractures, joints, or other openings which permit downward penetration of tree roots.

Alluvial soils, as understood by soil scientists, are soils developed from recently deposited alluvium. Because of their youth, alluvial soils show little or no profile development; the soil characteristics are determined primarily by the chemical composition and texture of the alluvium, and not by climate and vegetation. Soils with well-developed profiles which have developed from older alluvial deposits are not regarded as alluvial soils but are referred to the appropriate zonal or intrazonal group.

Alluvial soils occur in all parts of the United States but occupy particularly extensive areas along the lower course of the Mississippi River and its larger tributaries. Smaller areas, in the form of strips of varying width, are found along other waterways.

In humid and subhumid regions alluvial soils originally supported luxuriant forest vegetation. Extensive areas have been

cleared for agricultural production, but many others, because of poor drainage and susceptibility to inundation during periods of high water, have remained in forest. In texture, alluvial soils vary from sands to clays. The inherent fertility is usually high; in fact, these soils are among the most productive in the world.

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Chapter 12

FOREST SOIL CLASSIFICATION

Classification of soils involves their arrangement or distribution in appropriate, mutually exclusive categories. The purposes of soil classifications are similar to those of classifications of natural objects in general, namely, to facilitate organization of facts concerning the objects and to discover order. In the words of Miller (1851: 80) classification is a principle, "... which we find pervading all science — which gives to each of the many cells of recollection its appropriate facts — and without which all knowledge would exist as a disorderly and shapeless mass, too huge for the memory to grasp, and too heterogeneous for the understanding to employ."

All systems of classification of natural objects are more or less provisory, because they are based on incomplete knowledge of the objects themselves. Consequently, classifications must be modified from time to time as new facts are established.

DEVELOPMENT OF SYSTEMS OF SOIL CLASSIFICATION

As indicated in Chapter 3, the earliest classifications of soils were based on geology. Subsequently, as a result of the researches of Russian soil scientists, schemes based largely on climate were developed. Both of these types of classification were based on factors which influenced soil development and soil characteristics rather than on characteristics of the soils themselves. It is not surprising that both bases of classification were found inadequate when applied to a country having the variety of parent materials, climatic conditions, and vegetation types found in America. For details concerning the early classification systems Glinka (1927), Ramann (1928), Joffe (1936), and Baldwin *et al.* (1938) should be consulted.

The classification currently employed in America by the Division of Soil Survey of the Bureau of Plant Industry, Soils, and Agricultural Engineering is a natural system based on the

characteristics of the soil itself. All features which are believed to influence the utilization of soil for agricultural purposes are considered in the field classification. This system, although not entirely satisfactory for the classification of forest soils, merits attention, because it is the basis of practically all soil surveys which have been conducted in America. Much of the following information concerning this system has been drawn from the reports of Marbut (1935), Kellogg (1937), and Baldwin *et al.* (1938).

LOWER CATEGORIES OF CLASSIFICATION

At the present time the Division of Soil Survey recognizes three soil categories or units in the field: (1) series, (2) type, and (3) phase. Of these categories the phase is the lowest, and the series the highest.

A *soil series* is defined as a group of soils having genetic horizons similar in differentiating characteristics and arrangement in the profile, excluding the texture of the surface soil, and developed from similar parent materials. In a given series texture of the material of the A horizon may vary considerably in different soils, but the texture of the lower horizons in different soils is essentially similar. Furthermore, all soils included in a series have essentially similar horizon characteristics, that is, thickness, color, structure, content of carbonates and other salts, reaction, and content of humus.

Two facts concerning the characteristics which form the basis for recognizing soil series should be appreciated. First, the allowable variation within a series is not susceptible of precise definition; and, second, the characteristics which form the basis for differentiation are those which lend themselves rather readily to mapping. Kellogg (1937: 90) has indicated that generally the differentiating characters are "... those observable and mappable characteristics which are known to have, or are likely to have, significance in the growth of native or crop plants. ... " A series is given the name of a place near the point where it was first defined, for example, Merrimac, Olympic, Holyoke, Au Train. Several thousand soil series have already been recognized in America, and as information develops many additional series will be defined.

A *soil type* is a group of soils having genetic horizons similar

in differentiating characteristics and arrangement in the profile, including the texture of the surface soil, and developed from similar parent material. Although a series may include members whose surface soil (the A horizon or, very roughly, the upper 6-8 in. of mineral soil material) varies considerably in texture, that is, from a sandy loam to a loam, no such variation is allowed in a type. Soil-type names are formed by combining the series name and the textural class name, for example, Au Train loamy sand, Cecil sandy loam. Appropriate qualifying terms, such as stony or gravelly, are included in the type name when necessary for proper characterization of the soil. The number of textural classes included in a series varies, depending on the maturity of the soil. In well-developed soils the number does not exceed three, but in alluvial soils, which are immature, the number may be much greater.

A *phase* may be recognized within a soil type. This category includes members having characteristics which vary from those which are normal for the type. The characteristics most frequently made the basis for recognizing phases are degree of slope, stoniness, and accelerated erosion. Soil phases as currently recognized appear to have more significance for the agriculturist than for the forester.

HIGHER CATEGORIES OF CLASSIFICATION

The *order* is the highest category of classification. Included are three members: (1) zonal, (2) intrazonal, and (3) azonal soils. These categories, which are based on geographic relations, have been explained in Chapter 11 and will not be considered further here. *Suborders* are broad categories which include the *great soil groups*. The relation of the suborders to the great soil groups is indicated on the following pages. Members of the great soil groups have common internal characteristics.

One or more *families*, or categories of related soil series, may be included in each great soil group. The soil series, however, have not generally been assigned to families.

ARRANGEMENT OF THE HIGHER CATEGORIES OF CLASSIFICATION

The orders, suborders, and great soil groups may be arranged according to the following scheme (Baldwin *et al.*, 1938, and

others). Orders are indicated by Roman numerals, suborders by capital letters, and great soil groups by Arabic numerals. Families, series, and types, because of their large numbers, are not shown.

I. Zonal soils

A. Soils of the cold zone

1. Tundra soils

B. Light-colored soils of arid regions

2. Desert soils

3. Red desert soils

4. Sierozem soils

5. Brown soils

6. Reddish-brown soils

C. Moderately dark-colored soils of semiarid grassland regions

7. Chestnut soils

8. Reddish-chestnut soils

D. Dark-colored soils of subhumid to humid grassland regions

9. Chernozem soils

10. Prairie soils

11. Reddish prairie soils

E. Soils of the forest-grassland transition

12. Degraded chernozem soils

13. Noncalcic brown soils

F. Light-colored podzolized soils of the cool-temperate to temperate forest regions

14. Podzol soils

15. Brown podzolic soils

16. Gray-brown podzolic soils

G. Podzolic and lateritic soils of warm-temperate to tropical forest regions

17. Yellow podzolic soils

18. Red podzolic soils

19. Yellowish-brown lateritic soils

20. Reddish-brown lateritic soils

21. Laterite soils

H. Red lateritic soils of regions having a Mediterranean type of climate

22. Terra rossa

II. Intrazonal soils

A. Halomorphic soils

1. Solonchak soils

2. Solonetz soils

3. Soloth soils

B. Hydromorphic soils

4. Wiesenböden (meadow soils)
5. Alpine meadow soils
6. Bog soils
7. Half-bog soils
8. Planosols
9. Ground-water podzol soils
10. Ground-water laterite soils

C. Calomorphie soils

11. Brown forest soils (Braunerde)
12. Rendzina soils

III. Azonal soils

1. Lithosols
2. Alluvial soils
3. Dry sands

The general characteristics of the great soil groups have been considered in Chapter 11; their geographical occurrence is indicated in Figures 38 and 39.

CLASSIFICATION OF FOREST SOILS

As indicated earlier, the classification currently employed in the Division of Soil Survey is not an entirely satisfactory basis for classifying soils for forestry purposes. It is not surprising that this situation prevails. Most soil surveys have been directed primarily toward the classification of soils from the standpoint of their utilization for agriculture. In view of the fundamental differences between agricultural and silvicultural techniques and the differences in morphology and physiology of agricultural crop plants and forest trees, a single system of soil classification appears inadequate. Unfortunately, however, neither soil scientists nor foresters generally possess the information requisite to the formulation of a system really satisfactory for forestry purposes.

Through the years agronomists have accumulated a substantial body of quantitative data on soil and crop-plant relationships. Analogous information concerning the significance of various observable soil characteristics with respect to forest composition and growth is badly needed by foresters. Soil characteristics which are useful in the differentiation of agricultural soils may be of little or no significance from the standpoint of forest production. For example, slope and stoniness are highly important

in determining the suitability of land for the use of agricultural machinery, but these factors are of much less importance in forestry. On the other hand, the nature of the strata underlying the parent material and the occurrence of a water-table several feet below the soil surface influence the development of deep-rooted trees but have little effect on shallow-rooted agricultural crops. Forest soils usually possess, whereas agricultural soils lack, layers of unincorporated organic matter. Available evidence indicates that the nature of the humus layers of forest soils influences tree development and that this feature is worthy of consideration in the scheme of classification.

Soil classifications and site classifications are not synonymous, for the reason that site is determined by the integrated influence of *all* the factors of the environment, of which soil, although highly important, is only one. Climatic (atmospheric) and biotic factors, as well as soil, contribute to the sum total of the conditions which determine site. A further category of factors, namely, physiographic, is commonly recognized, but physiographic factors involve mainly soil and climate. Too frequently differences in site quality have been attributed to differences in soils simply because the observer failed to appreciate the ecological importance of the climatic and biotic factors.

CLASSIFICATION WITH RESPECT TO FOREST COMPOSITION

It is too much to expect that any soil classification can be developed which will serve as a basis for accurately delimiting forest types. In the opinion of Sampson (1930), "Biotic competition, rate of succession, habitat range, preoccupation by approximately ecological equivalents, biotic history, topographic and climatic factors may frequently be paramount over the differences in soil characteristics selected by the U. S. Bureau of Soils to delimit soil types." It has frequently been found, particularly in regions where natural conditions have been drastically modified, that a given soil type may support different forest types and that different soil types may support the same forest type (Hicock *et al.*, 1931; Roe, 1935; Turner, 1940).

Promising results have been obtained, however, in numerous regions where individual soil types were grouped together. Veatch (1928, 1932, 1933) has demonstrated a relationship between soil groups in Michigan and vegetation cover. In 1933 he

recognized eleven major land types and indicated the vegetation characteristic of each type. Westveld (1933), working in the Upper Peninsula of Michigan, also found a relationship between groups of soil types and forest composition. A similar relationship was also reported by Roe (1935). During the course of the forest survey in the South no relationship was found between forest types and soil types in the loessial region of northern Mississippi. However, it is now recognized as significant that upland hardwood forest is peculiar to the soil group in this region, which comprises soils of the Memphis, Loring, Grenada, and Lexington series (Turner, 1940). Turner stated that in the Little Tallahatchie River flood-control survey it was found that forest subtypes, dominated by specific groups of species of the general upland hardwood forest, are more or less peculiar to each of the types of loessial soil.

From time to time general relationships between indicator plants and forest-stand composition have been reported. Heimbürger (1934, 1941) found relationships of this kind in both the Adirondack Mountains in New York and in the Lake Edward Experimental Area in Canada. Often, however, no clear differences in lesser vegetation can be recognized in different forest types.

Attempts to relate forest composition to single physical, chemical, and biological characteristics of soils have been largely unsuccessful. This situation is to be expected, because forest composition is determined by the joint action of all habitat factors rather than by the influence of a single factor.

CLASSIFICATION WITH RESPECT TO GROWTH AND YIELD OF FOREST STANDS

The classification of soils from the standpoint of their capacity to produce timber stands has received increasing attention in recent years. Many different criteria have been used with varying degrees of success. General experience has shown that a single soil characteristic seldom forms as reliable a basis for evaluation of site quality as does a combination or group of characteristics (Erdmann, 1924; Wiedemann, 1934: 18; Aaltonen, 1937; Turner, 1937). It has been observed repeatedly that the correlation between soil categories and forest-site quality improves as the classification criteria are broadened to reflect as

completely as possible the entire complex of ecologically important soil characteristics. Perfect correlations between soil categories and site quality need not be expected because, as has been pointed out, soil conditions alone are not decisive.

Many attempts have been made to correlate individual soil types, as recognized by the Division of Soil Survey, with site quality. Usually the results have been unsatisfactory (Hicock *et al.*, 1931; Auten, 1933, 1936). On the other hand, various investigators have demonstrated that soil complexes, comprising groups of ecologically similar soil types, afford a useful basis for evaluating site quality.

Westveld (1933) found a relationship between several soil groups and site quality in northern Michigan. Twenty-three soil types were arranged in 7 soil groups, based on similarity of surface soil and subsoil characteristics. Three site index classes were represented, namely, 70, 80, and 90; the corresponding volumes per acre were 20,100, 23,800, and 29,000 ft. board measure, respectively. In general, there was little variation in site index on different areas of a given soil type. When variations in site index on a given soil type occurred, they were usually associated with variations from the typical soil condition. Five of the 7 soil groups recognized by Westveld are the following:

1. Loams, sandy loams, and loamy sands with a yellow sand substratum 15-30 in. below the surface.
2. Loams and sandy loams with a sandy clay till or drift substratum 25-30 in. below the surface.
3. Silt loams and loams with clay or sandy clay subsoils.
4. Silt loam and loam with open coarse sand, gravel, and cobbles below 40 in.
5. Shallow soils resting on bedrock.

The productivity of soils in groups 2, 3, and 4 was high when they were deep and reasonably free from rock; the productivity of soils in group 1 was intermediate; in group 5 it was intermediate to low. Northern hardwoods occupied the areas investigated.

In a more extensive investigation in the Lake States, Roe (1935) grouped some 400 soil types into 14 categories, based primarily on moisture relationships, as determined by the texture of the surface layers and the nature of the subsoil, in conjunction with the relative depth of the water-table. Two climatic regions

were recognized, namely, one characterized by northern hardwood forest and one characterized by white spruce-balsam fir. Texture classes employed were as follows: sands, light loams and loamy sands, loams and heavy loams, "variable," "mucky," and "peats and mucks." It is hardly necessary to point out that the last three categories are not, in reality, textural classes at all. Subsoils were classified as rocky, sandy, or clayey. Moisture relationships were designated as dry, fresh, moist, intermittently wet, wet, and saturated. Differences in the site quality of the various soil groups were established.

In 1938 Kittredge reported on an investigation concerning the relationship of growth of aspen stands to soils in Minnesota and Wisconsin. Fifty-four soil types were represented, and 4 site-quality classes were distinguished. Growth was correlated with various soil conditions as follows:

Basis of Classification	Correlation Ratio
Soil-texture classes	0.573
Surface formations	0.640
Soil-texture classes and surface formations	0.699
Soil-profile groups	0.788

The number of soil-texture classes recognized was 8. Twelve surface formation, or geological formation, groups, for example, outwash, sandy moraine, till, and peat, were distinguished. Soil types were arranged in 12 profile groups on the basis of the nature of the *C* horizon (texture, water-holding capacity, and lime content), development of the *A*₂ horizon, general moisture relations, texture of the surface soil, and presence or absence of glei horizons. It will be noted that the correlation between growth of aspen and soil conditions improved as the classification criteria were broadened.

Turner (1937) established a relation between the growth of loblolly and shortleaf pine in the coastal plain region of Arkansas and so-called site-soil complexes, as expressed in phases of specific soil types or in certain soil-profile patterns. Site index for loblolly pine was found to vary from 110 ft. on Ochlockonee silt loam and fine sandy loam to 60 ft. on Prescott very fine sandy loam. Turner reported that certain soil types, or phases of types,

in the coastal plain of Arkansas are adapted to the requirements of loblolly pine only, others to both shortleaf and loblolly pine, and some to shortleaf pine only.

Numerous schemes of soil classification which, in contrast to the foregoing, are more or less independent of the soil types as recognized by the Division of Soil Survey have been developed for forestry purposes. Wilde and Scholz (1934) classified the soils of the Upper Peninsula Experimental Forest in Michigan into five categories:

1. Slightly podzolized soils (dark forest loam, mull loam). Productivity high.
2. Podzol soils (strongly leached and cemented loam). Productivity medium to high.
3. Swampy podzol (mottled loam, glei soil). Productivity low.
4. Muck (peaty-clay soil, sapropel soil). Productivity low.
5. Woody peat. Productivity very low.

A classification of glei soils was proposed by Wilde in 1940. The following three categories were recognized:

1. Alpha-glei; glei within about 1 ft. of the surface. Principal species are balsam fir, white spruce, and some black ash and red maple.
2. Beta-glei; glei 2-3 ft. from the surface. Principal species are sugar maple, rock elm, red maple, and some basswood, yellow birch, balsam fir, and white spruce.
3. Gamma-glei; glei 5-7 ft. from the surface; mottling may be observed at a depth of about 4 ft. Principal species are sugar maple, basswood, and some eastern white pine.

Yields of fully stocked stands on these soils in the zone of podzols in Wisconsin are indicated as being 2,200, 3,500, and 4,800 cu. ft. per acre, respectively. Values are based on a rotation age of 100 years.

Relationships have been shown to exist between various other soil characteristics and forest growth. Valmari (1921) regarded soil analysis as the only exact basis for classification of soils into productivity classes. He showed that site quality in Finland improved with increase in nitrogen and calcium in the upper soil layers. The findings of Valmari were confirmed by the work of

Ilvessalo (1923) and Aaltonen (1926). Haig (1929) found the values for both colloidal content and silt plus clay of Connecticut soils correlated with site quality as indicated by the height growth in young Norway pine plantations. Hicock *et al.* (1931) also investigated the relation of soil characteristics to growth of Norway pine in Connecticut. The total nitrogen content of the A_1 and A_2 horizons showed a better correlation with site index than any other factor analyzed. A fairly high correlation was also noted between the moisture equivalent of the A_1 and A_2 horizons and site index. In 1935 Coile examined the soil profiles in 48 even-aged shortleaf pine stands in the Duke Forest. No single soil characteristic of any one horizon was highly correlated with site index in all the stands investigated. Coile found that site index was influenced by the texture of the B horizon and the depth below the surface at which the B horizon occurred. A texture-depth index,

Content of silt + clay in the B horizon, in per cent

Depth of the B horizon below the surface, in inches

which proved useful in establishing site quality, was developed. Improvement in site quality with an increase in the texture-depth index from 1 to 5 was rapid. Soils having a texture-depth index of 5 appeared to have the highest site index; above 5 there was a decrease in site quality with an increase in texture-depth index.

Auten (1936, 1945a) found that growth of black locust and eastern black walnut in the Central States was closely correlated with the properties of the subsoil which influence drainage and aeration, namely, plasticity, structure, and compactness. Soils that have a compact, plastic subsoil at a depth of less than 8-14 in. are poor sites for both species. The color of the subsoil is a useful characteristic in judging drainage conditions.

Auten (1937a, 1937b, 1945b) also investigated the site requirements of tulip tree. It was established that the depth of penetration of organic matter in the surface soil horizon affords a useful criterion of the site index. If the depth of penetration (A_1 horizon) is less than 1 in., tulip tree does not show satisfactory development; average or better sites for this species usually have an A_1 horizon 3 or more in. thick. The depth to tight subsoil is an even better criterion of site index for tulip tree.

The presence of a tight subsoil less than 24 in. below the surface results in poorer than average sites. Auten also reported that the topographic position of the soil body influences site index. The best growth was observed in sheltered coves and the poorest on ridges. Within the range of soils investigated no correlation could be shown between the growth of tulip tree and the amounts of available nutrients in the soil. Turner (1938) reported relationships between certain soil features and the growth of short-leaf and loblolly pine in the coastal plain of Arkansas. He established a definite correlation between height growth and the depth of the B_1 horizon; a correlation was also established between growth and the content of clay in the B_2 horizon. Growth increased as the clay content of the B_2 horizon decreased. Turner also found that growth decreased as the degree of slope increased, especially on slopes of more than 3 per cent.

Classification of forest sites on the basis of the lesser vegetation (indicator plants) has received, and is still receiving, much attention by foresters. The historical development of the concept has been reviewed in detail by Heimburger (1934). Finnish foresters have been particularly active in developing and applying in practice the theory of site types. In 1926 Cajander defined the Finnish forest (site) type as follows:

... all those stands are referred to the same forest type the vegetation of which, at or near the time of maturity of the stands and provided the stands are normally stocked, is characterized by a more or less identical floristic composition and by an identical ecologico-biological nature, as well as all those stands the vegetation of which differs from that defined above only in those respects which — being expressions of differences due to age, fellings, etc. — have to be regarded as merely accidental and ephemeral or at any rate as only temporary. Permanent differences call forth a new forest type in cases where they are sufficiently well-marked, or a sub-type in cases where they are less essential, but, nevertheless, noticeable. In a forest type, therefore, as a rule, only those primary — climatic and edaphic — factors of the locality are reflected — which factors may be assumed to remain active, even when the locality is laid bare of all plants.

Although the theory of indicator types, as applied to the classification of forest sites, has frequently been criticized, as will be discussed later, it has been successfully used in many parts of

the world (Valmari, 1921; Ilvessalo, 1923; Aaltonen, 1929; Bornebusch, 1923-1926, 1931; Wiedemann, 1934; Heimbürger, 1934; Kittredge, 1938; Sisam, 1938; Ray, 1941). Heimbürger found a general relation between the growth of red spruce and site types in the Adirondacks. Kittredge recognized 16 plant-indicator groups in his investigation of aspen communities in Minnesota and Wisconsin. The correlation ratio of plant-indicator groups with the mean site index of aspen was 0.761. This correlation was superior to that obtained with any soil groupings except soil-profile groups (0.788). Sisam recognized four site types for aspen in the Petawawa Forest Reserve in Canada. These site types, arranged in order of decreasing productivity, were *Aralia*, *Trillium*, *Aster-Corylus*, and *Maianthemum-Corylus*. The average heights attained by 60-year-old trees 12 in. in diameter in each site type were 94, 91, 79, and 76 ft., respectively. More recent work by Ray in red spruce and balsam fir stands in the Lake Edward Forest Experimental Area in Canada resulted in recognition of three site types, namely *Cornus*, *Oxalis-Cornus*, and *Viburnum-Oxalis*, which included over 90 per cent of the area. Ray concluded that site types are accurate expressions of site quality.

The use of the lesser vegetation as a basis for classifying forest sites has been criticized by various investigators (Burger, 1926; Hesselman, 1926: 538-542; Coile, 1938). The following criticisms are most frequently made:

1. Growth of forest stands varies, sometimes materially, on a given site type.
2. Forest composition, stand density, competition, and past history of the areas influence the lesser vegetation.
3. The lesser vegetation, which is relatively shallow-rooted, does not reflect conditions in the subsoil horizons, although these conditions influence the growth of trees and hence site quality.

Although some of the classifications of forest sites on the basis of indicator plants appear highly subjective (for example, in the *Oxalis-Myrtillus* type the characteristic plant *Oxalis acetosella* may be lacking), it should be recognized that in a general way relationships exist between ground vegetation and soil productivity. Indicator vegetation should be a very useful adjunct to other methods of evaluating site quality. However, before full

use can be made of ground vegetation as an indicator of site conditions, it will be necessary to acquire substantially more knowledge of the ecology and successional relations of herbaceous and shrubby plants than most foresters now possess.

CLASSIFICATION WITH RESPECT TO REFORESTATION

It is becoming increasingly evident that success of a reforestation program is contingent on proper evaluation of the sites to be planted in terms of survival and growth of the planting stock. Evaluation of site permits concentration of effort on areas where natural conditions are most favorable and aids in intelligent selection of species to be planted.

Diebold (1935) grouped 19 soil types in New York into 9 categories. The basis for grouping was the lime content in the entire soil profile, the degree of drainage, and the depth, structure, and consistency of the subsoil. He recognized drainage conditions as particularly important in evaluating soils from the standpoint of reforestation. Survival on wet soils was low because of frost heaving; growth on these soils was also poor. Le Barron *et al.* (1938) investigated the influence of surface-soil texture on survival of planted Norway pine in the Superior National Forest in northeastern Minnesota. Three broad textural classes were recognized: (1) light soils with a silt and clay content of less than 15 per cent, (2) medium soils with a silt and clay content of 15-30 per cent, and (3) heavy soils with a silt and clay content of more than 30 per cent. The highest survival in the areas planted in the spring occurred on the heavier soils, but survival of stock planted in the fall was highest on the light and medium-textured soils. This finding resulted in the practical recommendation that, as far as possible, planting should be done in the spring and that such fall planting as was necessary should be confined to the lighter soils.

A useful site classification for reforestation in the National Forests of Wisconsin has been developed by Stoeckeler and Limstrom (1942). Four major factors are considered in this classification: soil texture, depth to the water-table, presence or absence of an overstory, and density of the ground cover. The first two of these are soil factors, and the third and fourth involve the vegetation present on the site. Five texture classes,

based on the average silt-plus-clay content of the *A* and *B* horizons, are recognized: 0-10 per cent, light sands; 10-20 per cent, sands and loamy sands; 20-30 per cent, light sandy loams; 30-50 per cent, sandy loams and loams; more than 50 per cent, heavy soils.

Two situations are recognized with respect to water-table relations: (1) beneficial water-table present (2-6 ft. below the surface in midsummer on planting sites and 2-4 ft. below the surface on sites for direct seeding of pine), or (2) beneficial water-table absent, that is, occurring at depths greater than those noted in (1). Species and age of stock are recommended for each of 80 combinations of soil and cover conditions.

Development of black locust on old field soils in Lauderdale County, Mississippi, was investigated by Roberts (1939). It was found that the depth of the surface soil material down to the compact sandy clay or clay layer had a very important bearing on height growth. After four growing seasons the following relation was found between depth of the surface soil and average height of the trees:

Depth of the Surface Soil, inches	Average Height, feet
0.5	1.95
5.0	3.65
10.0	5.41
20.0	8.43
30.0	10.83
38.0	12.16

The conclusion was reached that the effective use of black locust in erosion control is limited to situations having considerable surface soil.

Investigations in eastern Tennessee have recently demonstrated remarkable differences in survival in old field plantations on different soils and aspects (Minckler, 1941*a*, 1941*b*). Three soil categories were recognized as follows:

1. Soils derived from dolomite limestone (Clarksville and Fullerton series)
 2. Soils derived from limestone (Dewey and Talbott series)
 3. Soils derived from shale (Montevallo and Armuchee series)
- Northerly and southerly slope aspects were distinguished. At

the end of the first year after planting the percentage survival of five species (shortleaf pine, eastern white pine, tulip tree, eastern black walnut, white ash) grouped together was as follows:

Year	Aspect	Soil Category		
		<i>Dolomite</i>	<i>Limestone</i>	<i>Shale</i>
1938 (wet)	Northerly	94.8	85.4	87.6
	Southerly	92.8	92.7	88.0
1939 (dry)	Northerly	91.1	89.4	79.7
	Southerly	89.3	80.3	49.1

The survival of shortleaf pine was best on soils derived from dolomite limestone on both slope aspects and on limestone soils on southerly aspects. A dense cover of competing vegetation characterized the limestone and shale soils on slopes having a northerly aspect. Moisture relations were unfavorable in shale soils on south slopes, and survival was low. Survival of eastern black walnut was relatively poor only on south slopes with dolomite soil. White ash survival was equally good on all soils and aspects.

Utilizing these data, Minckler (1941b) developed a practical planting chart, which promises to contribute materially to successful reforestation of old field areas in the Appalachian Valley in eastern Tennessee. The chart is based on slope aspect, soil depth, topographic position, soil-profile characteristics (principally physical), and density of native vegetation. Specific recommendations of species to be utilized in reforestation of the various sites are provided. More recent investigations by Minckler (1943) have contributed additional evidence of the important influence that the depth of the *A* horizon and the nature of the *B* horizon exert on forest plantations. The consistency of the *B* horizon was highly important for all species except shortleaf pine. During years of low rainfall, mortality was greatest and growth poorest on soils having a stiff *B* horizon. As would be expected, growth was best on soils with the deepest *A* horizons.

Indicator vegetation has been used to a limited extent to evaluate planting sites. Stewart (1935) recognized four soil-fertility levels in New England old fields on the basis of the species composition of the native vegetation. Similarly, Hursh and Crafton (1935) reported that the moisture and nutrient conditions of soils of the Piedmont and lower mountain slopes of the Carolinas could be judged by the plants that appeared on the

fields during the first 10 years after their abandonment. These investigators also recognized that the species composition of the native vegetation undergoes change with successional development. Four classes of sites were characterized on the basis of composition of the native vegetation:

1. Soils with minimum amounts of moisture and nutrients (the poorest sites).
2. Soils with low amounts of moisture and nutrients (poor sites).
3. Soils with medium amounts of moisture and nutrients (fair sites).
4. Soils with relatively high amounts of moisture and nutrients (good sites).

General recommendations as to choice of species for planting on each site were given.

Much additional research is needed to establish relations between observable soil characteristics and survival and growth of plantation stands. Only when such relations have been determined will it be possible to work out really satisfactory schemes of soil classification for reforestation purposes. The influence of the chemical, physical, and biological properties of the soil profile on tree development all merit careful consideration. Criteria which prove satisfactory for classification of soils in one region may be quite inadequate in another region of different soils and climate.

IMPORTANT CRITERIA FOR CLASSIFICATION OF FOREST SOILS

The criteria which are used for classifying forest soils vary with the objective of the classification. Criteria which are important from the standpoint of reforestation or establishment of natural reproduction may be relatively less significant from the standpoint of subsequent rate of growth and yield. Pearson (1931: 125) has pointed out that as a rule conifers in the Southwest reproduce with difficulty in depressions having fine-textured soil. When these situations are adequately stocked, however, they produce the heaviest stands of timber. In 1924 Veatch recognized that the soil conditions most important from the standpoint of seedling survival, rate of growth, size of mature trees, density of stands, and nature of root systems differed.

The emphasis to be placed on various criteria of classification

vary in different climatic and topographic regions and in different soils. For example, aspect is probably most important in regions having a cold or dry climate. Degree of slope merits more consideration where erosion is severe than where it is negligible.

Two general categories of soil characteristics may be recognized: (1) those which are relatively permanent, and (2) those which are subject to appreciable change during relatively short periods of time.

1. Characters which are relatively permanent
 - a. Thickness and texture of the A horizon
 - b. Thickness, texture, structure, and consistency of the B horizon
 - c. Texture and structure of the C horizon
 - d. Nature of the underlying strata with respect to permeability to water and roots
 - e. Nature of the parent material with respect to content of calcium (calcareous, noncalcareous)
 - f. Position of the water-table
 - g. Content of rock in the soil
 - h. Slope (degree and topographic position)
 - i. Aspect
2. Characters which are subject to appreciable change during relatively short periods of time
 - a. Humus layer type
 - b. Nitrogen content of the surface soil layers
 - c. Structure of the surface soil layers

These soil characteristics are believed to include those most important from the standpoint of forest production. Probably no one classification scheme will employ all the characteristics listed, but every classification will utilize some of them.

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Chapter 13

SOIL EROSION AND FOREST-SOIL DETERIORATION

Soil erosion is a serious problem in America and in many other parts of the world. This fact has been emphasized repeatedly by geologists, soil scientists, agronomists, graziers, and foresters, especially during the past 20 years (Marsh, 1907; Glenn, 1911; McGee, 1911; Free, 1911; Sampson and Weyl, 1918; Zon, 1927; Bennett and Chapline, 1928; Utz *et al.*, 1938; Bennett and Lowdermilk, 1938; and many others). Some conception of the magnitude of the problem may be gained from the fact that of the total land area of the United States, excluding mountains, mesas, and "bad lands," approximately 282,000,000 acres have been either ruined or severely damaged by erosion. Equally impressive is the fact that from one-quarter to three-quarters of the topsoil has been removed from an additional 775,000,000 acres. The importance of soil erosion is further evidenced by the voluminous literature on the subject (Williams *et al.*, 1937; Gaines *et al.*, 1938).

The discussion in this chapter will not be restricted to erosion of forest soils; brief consideration will also be given to erosion on agricultural and grazing lands, since to a very large extent the erosion problem concerns lands in these two categories. On first thought it may appear that erosion on nonforested areas is of no concern to foresters. On further consideration, however, it becomes evident that they have a very real interest in the matter. Some of the most difficult reforestation problems confronting foresters today are the direct outcome of accelerated erosion on mismanaged agricultural and grazing lands. It is also a fact that economic and other considerations occasionally make it desirable that open areas located within forest units be continued in either agricultural or grazing use. Intelligent administration of these areas requires that the forester in charge have at least a working knowledge of the factors affecting erosion and the principles of erosion control as they apply to nonforested lands. Without this knowledge the forester is not fully equipped to discharge his duties as a land manager.

Soil erosion is the process of removal of soil material by running water, wind, or gravitational creep. The process is natural; it is of universal occurrence and has operated throughout the past, even prior to the coming of man.

Two categories of soil erosion are commonly recognized. *Normal erosion* is characteristic of the land surface in its natural environment. It is sometimes referred to as geological erosion. *Accelerated erosion* is erosion in excess of the normal; it results, for the most part, from changes in the nature and density of native vegetation due to human activity. Normal erosion usually proceeds at a very slow rate in forest regions; in general, it is beneficial from the standpoint of maintenance of soils in a productive condition, since it proceeds at a pace no more rapid than that at which new soil is formed from the parent material. If land is so flat that normal erosion does not occur, the soils ultimately reach a stage of senility. The large group of intrazonal soils known as planosols are a good example of the unfavorable conditions that result when normal erosion is lacking. Planosols have strongly leached, acid A horizons and extremely heavy-textured B horizons; fertility and drainage of these soils are commonly poor. Accelerated erosion, on the other hand, is rapid and is invariably harmful, because fertile topsoil is removed at a rate greatly exceeding that of soil formation.

Efforts to control or reduce normal erosion are seldom justified in forest or range management but may occasionally be warranted by agronomists to protect adjacent valuable land. Prevention or, where this is not feasible, reduction of accelerated erosion, however, is not only justified but also is essential to the maintenance of soil productivity and rational land use. The ramifications of the problem of accelerated erosion are many. They include not only the losses of soil from the land but also the effects on floods, navigation, water-power development, production of potable water, water conservation, and wild life. The discussion which follows, therefore, is concerned almost exclusively with accelerated erosion.

TYPES OF SOIL EROSION

The natural agencies chiefly responsible for accelerated erosion are water and wind. Water erosion is active in all regions, arid as well as humid. Erosion by wind is most active in arid or semiarid

regions but is by no means unknown in humid regions. Erosion controlled directly by gravity occurs in sections of pronounced topographic relief. For the most part erosion due to the influence of gravity is normal.

EROSION BY WATER

Erosion by flowing water involves hydraulic action, abrasion, solution, and transportation. Hydraulic action involves the force inherent in the flow of water which is able to lift and move solid particles. Abrasion results from the friction of particles carried in suspension or moved by traction and particles in the bed or sides of the channel. In addition, there is some friction between the particles in suspension as a result of differences in their direction and rate of movement. Solution occurs whenever water comes in contact with soil material and materially aids erosion. Transportation of soil particles is the logical outcome of hydraulic action, abrasion, and solution. Operation of these processes always involves a change in position of the materials concerned. Suspension and traction both play a part in transportation by water; particles that are moved by traction roll, slide, or make small jumps along the bed of the channel.

Velocity of flow has an important bearing on the erosive power of water. The weight of the largest particles which can be moved by a stream varies as the sixth power of the "bed" velocity (Rubey, 1938). The unit width load (capacity) in a stream free to pick up sand and gravel as its rate of flow is increased is stated by Rubey to vary roughly as the third power of the "bed" velocity. With an increase of velocity an increase in abrasive power is also observed; it varies between the square and sixth power of the velocity.

Sheet erosion is a term applied to the removal of a more or less uniform layer of material from the soil surface. The amount of soil removed during any one storm or in any one year may be relatively small, but over a period of years the fertile A horizon is washed away. The general contour of the land surface is not altered by sheet erosion; for this reason the insidious effects may go unnoticed until most of the surface soil has been removed. Probably sheet erosion does more damage to agricultural lands than does the more spectacular rill and gully erosion. Sheet

erosion is unimportant in forests having more or less continuous layers of unincorporated organic matter.

Rill erosion (fingering, shoe-string gullying, or incipient gully-ing) is applied to the type of erosion which results in the formation of small channels in the land surface. The channels are relatively shallow and in agricultural land are obliterated by tillage. Rill erosion is a stage intermediate between sheet and gully erosion. In forested areas rill erosion may occur along skid roads and on burns.

Gully erosion produces large channels, or gullies, which are not obliterated by ordinary tillage. These large channels, which may be more than 100 ft. deep, are developed in situations where



FIG. 45. This large gully in Benton County, Mississippi, is about 3000 ft. long, 276 ft. wide, and 76 ft. deep. Photograph by United States Forest Service.

the concentrated run-off has attained sufficient volume and velocity to cut deeply into the soil body (Figure 45). Gullies may have their beginning in the small channels caused by rill erosion, or they may form in furrows, skid roads, wagon roads, trails, or natural depressions. The shape that a gully assumes is influenced by the nature of the soil and subsoil. If the material is easily eroded, the side walls tend to be nearly vertical, as they frequently are in loessial soils and in the alluvial soils in the West

(Figure 46). On the other hand, a subsoil and underlying stratum resistant to erosion favor development of gullies with rather gently sloping sides. As a rule, gullies carry water only dur-

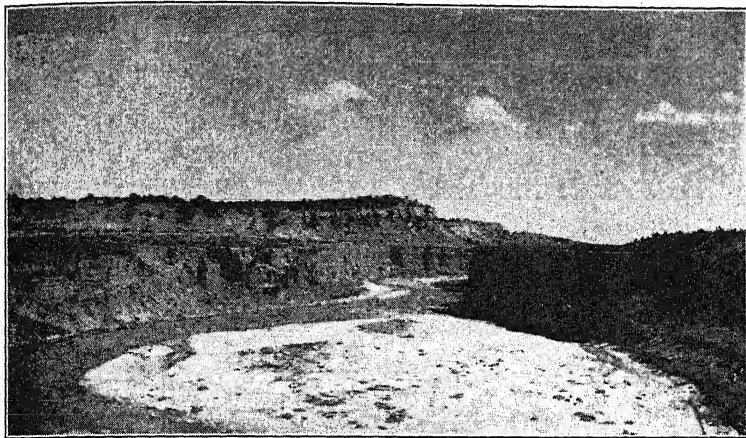


FIG. 46. The nearly vertical side walls of this gully, Rio Puerco arroyo, New Mexico, are about 50 ft. high. Photograph by H. H. Chapman.

ing or immediately after rains or after the melting of snow. Gullying proceeds by waterfall erosion at the gully head, by channel erosion, and by erosion caused by alternate freezing and thawing. All three processes are commonly active in the same gully.

Bank-cutting is a term used in referring to the erosion of banks of streams and rivers. The thread of greatest velocity in a stream is on the convex side of the current and is consequently directed against the concave bank. Undercutting occurs, with the result that the bank is gradually eroded away. Along exposed lake and ocean shores severe erosion sometimes results from wave action.

EROSION BY WIND

Because of its greater tenuity, air has a much lower transporting power than water. As a rule, only the relatively small-sized soil particles are moved by wind. Consequently, when wind attacks heterogeneous material, size sorting occurs; the finer particles are removed, and the coarser are left behind. The actual size of particles which can be transported, or the competence of

the wind, depends on the shape and density of the particles and on the velocity of the wind. Transport by wind may be either up hill or down, and the materials may be moved either by traction or in suspension.

Erosion of soil by wind assumes important proportions in dry regions where the topography is flat or gently undulating and where the natural vegetation is scanty or has been drastically disturbed. Areas of this kind are found throughout the Great Plains region, particularly in the so-called dust bowl in southeastern Colorado, southwestern Kansas, and the panhandle areas in Oklahoma and Texas. Many of the soils of arid and semiarid regions have a single-grain structure which is natural or which is induced as a result of agricultural or grazing use. Aggregates, if any, tend to be small. Soils of this kind are very susceptible to wind erosion when dry.

In forest regions important wind erosion is usually limited to areas of sandy soil near the shores of oceans or large lakes where wind velocities are high. Examples are known at various points along the Atlantic and Pacific coasts and along the eastern shore of Lake Michigan. If not covered with forest or other vegetation, the marine and lake sands are readily shifted about by the winds to form dunes. Such sands contain very little organic or inorganic colloidal material and consequently possess practically no cohesion when dry.

EROSION BY GRAVITATIONAL FORCES

Two types of mass movements of soil which are controlled directly by gravity may be recognized: (1) slow mass movement, or *creep*, and (2) rapid mass movement, or *landslides*. Soil creep occurs on steep slopes; the movement is so slow as to be scarcely perceptible. The entire mass of soil moves as a unit, so that no breaks are evident in the surface. A high content of water in the soil mass favors creep by adding to the weight and by serving as a lubricant, thus reducing friction.

Landslides usually occur in regions of precipitous slopes. Sometimes great masses of soil and rocks suddenly rush down the slopes, leaving behind paths of destruction which persist for many years. The debris piles up at the base of the slope, occasionally damming a stream. At other times the movement may be intermittent, following very heavy rains or occurring

when the soil thaws in the spring, with displacement of only a few feet at any one time. If the land bears forest, most of the trees may continue to grow, although some will be forced into leaning positions or uprooted. Intermittent mass movements of soil have been observed repeatedly in eastern Ohio, West Virginia, and southwestern Pennsylvania. They are also known to occur in the Washakie National Forest in Wyoming.

Rapid mass movements are likely to occur where the soils have been derived from slickensided granitic or gneissose rocks. Glenn (1911) has pointed out that the slickensided surfaces are preserved in the soil and, when wet, become planes of easy movement. Landslides are also likely to occur in situations where there is a sharp plane of division between the soil and the underlying solid rocks. This condition exists where glacial till has been deposited as a mantle on relatively smooth rock surfaces and where weathering of the exfoliation type has occurred. It is believed that earthquakes are one of the agents responsible for initiating rapid mass movements of soil.

FACTORS AFFECTING WATER AND WIND EROSION

Various factors are known to influence erosion. They may be indicated as: (1) climate, (2) nature of the soil, (3) slope, (4) vegetation, and (5) cultivation and cropping practice. It should be recognized that the severity of erosion is usually dependent on all these factors and not on any one alone.

THE INFLUENCE OF CLIMATE

Of the climatic factors which influence water erosion, precipitation and temperature are most important; in wind erosion the factor of wind velocity is highly influential. Precipitation and wind movement determine the activity of the erosional agents and, together with temperature, have a profound influence on the nature of the soil and protective vegetation.

Blumenstock (1939) has indicated the importance of precipitation characteristics, such as rainfall intensity and duration and storm frequency. It is generally recognized that with increase in rainfall intensity there is a tendency for both run-off and erosion to increase. Visher (1941) found that during the decade 1929-39 there were 47 official records of rains of 10.0-14.9 in. of rain in 24 hours in the deep South. During the same period

there were only 8 comparable records in the upper South, 4 in the North, and 1 in the West. Water erosion in the South is especially severe, not alone because of the type of agriculture practiced, but also because torrential rains are exceptionally numerous. Intensity remaining the same, storms of long duration cause greater run-off and erosion than storms of short duration. Similarly, very frequent storms result in more erosion than do less frequent storms of equal intensity and duration. Of course, if rainfall is so infrequent as to cause drought conditions and a marked reduction in vegetation density, wind erosion may be severe. Furthermore, rainfall at very infrequent intervals, falling on poorly protected soil, does far more damage than would more frequent rains on lands supporting a dense mantle of vegetation.

The seasonal aspects of precipitation have a bearing on the erosion of agricultural soils. Precipitation coming at a time when the land is not protected by a crop cover may be particularly damaging. In northern regions erosion is inactive during much of the winter, when field soils are exposed, but in southern localities, where prolonged freezing does not occur, erosion is active throughout the winter. Erosion resulting from alternate freezing and thawing is prevalent in parts of the South. It is particularly noticeable on gully slopes having a southern exposure.

Climatic conditions favorable to wind erosion are low amounts of precipitation, low relative humidity of the atmosphere, and high wind velocities.

THE INFLUENCE OF SOIL CHARACTER

The amount of surface run-off is influenced to a marked degree by the infiltration capacity of the soil. In fact, surface run-off occurs only when rainfall intensity exceeds infiltration capacity. Soil conditions which have a bearing on the rate of infiltration of water have been discussed in Chapter 9. In general, all soil characteristics which contribute to a large noncapillary pore volume favor infiltration. In this connection it should be kept in mind that the rate of percolation of water through a soil is governed by the permeability of the horizon which is least pervious. This fact explains the relatively low infiltration capacity of soils having subsurface horizons of low permeability. Organic matter, both unincorporated and incorporated, greatly aids infiltra-

tion (Glenn, 1911; Lowdermilk, 1930; Veihmeyer, 1938; Duley and Kelly, 1941). The mantle of unincorporated organic matter in forests has much to do with the high infiltration capacity of forest soils. It is a matter of common observation that the rate of erosion increases as the surface soil is removed and subsoil low in organic matter is exposed.

The field capacity of soils is another factor influencing erosion. The higher this capacity is, the less water there is to run off, if infiltration is assumed to be good. Soils having a low field capacity are usually coarse in texture, poor in organic matter, and noncohesive, and may support sparse stands of vegetation. It should be recognized, however, that many sandy soils, although having low field capacity, are relatively nonerosive, because they are so porous that a high rate of infiltration is maintained.

Soil characteristics which influence the resistance of soil particles to dispersion and transport also have a bearing on erodibility. From the standpoint of texture it appears that soils containing high proportions of sand and silt are more readily dispersed than soils containing considerable clay. The cohesion of the clay particles tends to bind them together and prevent dispersion. Once dispersed, however, fine particles are much more readily transported than are coarse particles.

Erosion by both water and wind is less in well-aggregated soils than in those having small, unstable aggregates. A single-grain structure is distinctly unfavorable. The value of organic matter from the standpoint of soil structure is widely recognized. Browning (1938), for example, reported that the addition of organic matter to soils in West Virginia increased the number of large-sized aggregates. The soils to which organic matter was added originally contained only a small amount of material capable of binding the particles together in stable aggregates.

In an investigation of soil properties which influence erosion Middleton (1930) reported that the dispersion ratio, the ratio of colloid to moisture equivalent, and the erosion ratio were most important. The dispersion ratio is the ratio, expressed in percentage, of the suspension percentage (measured arbitrarily by determining the fraction of soil which remains suspended for a specified time when dispersed under certain conditions) to the percentage of the total silt and clay in the soil, as determined by mechanical analysis. It is evident that the dispersion ratio

bears a relationship to soil texture and the ease with which the particles are dispersed. Soils in which a high proportion of the silt and clay is easily dispersed have high dispersion ratios and tend to be more erosive than soils having low ratios. The ratio of colloid to moisture equivalent is regarded as an index of the permeability of the soil to water. It is thought that an increase of the colloid-moisture equivalent ratio indicates a decrease in the field capacity of the soil, which in turn may be expected to result in a higher rate of infiltration, with consequent reduction in run-off during storms. Soils having high ratios of colloid to moisture equivalent tend to be less erosive than soils having low ratios. The erosion ratio combines in a single expression the dispersion ratio and the colloid-moisture equivalent ratio. The combination was effected by dividing the dispersion ratio, which varies directly with erosivity, by the colloid-moisture equivalent ratio, which varies inversely with erosivity. The following tentative limiting values for the various measures were suggested for distinguishing erosive from nonerosive soils:

	Erosive Soils	Nonerosive Soils
Dispersion ratio	>15	<15
Ratio of colloid to moisture equivalent	< 1.5	> 1.5
Erosion ratio	>10	<10

Various investigators have subsequently confirmed the usefulness of the erosion ratio as an index of erodibility (Middleton *et al.*, 1934; Peele, 1938; Rost and Rowles, 1941). Rost and Rowles stated that in Minnesota cultivated soils with dispersion and erosion ratios of less than 19 may be expected to be resistant to erosion. They regarded soils with dispersion ratios of more than 19 and erosion ratios of more than 30 as susceptible to erosion. Lutz (1935) indicated that the degree of hydration of soil colloids influences erodibility. The colloid of Iredell sandy clay loam, an erosive soil of the Piedmont Plateau region of North Carolina, is hydrated, whereas the colloid of Davidson clay, a nonerosive soil, is nonhydrated. The Iredell sandy clay loam is readily dispersed, but the Davidson clay is not.

There appears to be a relation between the silica-sesquioxide ratio of soils and the ease with which they are eroded (Bennett, 1926; Middleton, 1930). Soils having a low silica-sesquioxide

ratio tend to be better aggregated and less easily dispersed than soils in which the ratio is high.

Soils containing substantial amounts of rock tend to be resistant to erosion. This fact was noted by Glenn (1911) in his studies in the southern Appalachian region and Monongahela basin.

THE INFLUENCE OF SLOPE

In general, run-off and erosion increase with increase in degree of slope (Conner *et al.*, 1930; Duley and Hays, 1932; Connaughton, 1935; Renner, 1936; Neal, 1938). Conner *et al.* reported the following average annual losses, based on a 3-year period, of soil per acre from plats planted to cotton in Texas:

Slope	Soil Loss, pounds per acre
Level	8,583
1 per cent	18,624
2 per cent	25,395
3 per cent	25,572

With increase in length of slope, erosion usually increases. Velocity of run-off is dependent on degree of slope, but the volume depends on the size of the drainage area. Long slopes tend to shed more water than short slopes, thus increasing erosion.

THE INFLUENCE OF VEGETATION

The agent most effective in preventing accelerated erosion in any region is the climax vegetation. Only slightly, if any, less effective are well-stocked plant communities representing stages of succession somewhat short of the climax.

In properly managed forest stands accelerated erosion is negligible. This fact has been demonstrated repeatedly in various parts of the world (Glenn, 1911; Zon, 1927; Munns *et al.*, 1938; Anon., 1940). Munns *et al.* (p. 611) stated that soil losses from forests are from 10 to 0.01 per cent of those from cultivated fields and frequently smaller than the losses from grassland. In a recent report of the Joint Committee on Forestry it was stated that during the 1931 flood in Mississippi 17 in. of the 21-in. rainstorm ran off experimental plots of cleared land, removing 34 tons of topsoil per acre. On adjoining forest land the run-off was only 0.1 in., with no loss of soil.

The principal reasons why a dense cover of vegetation reduces erosion are as follows:

1. Infiltration of water is favored. Soil porosity under vegetation tends to be high because of the relatively large amount of organic matter added annually. The unincorporated organic matter, which is characteristic of most forest soils, breaks the impact of rain drops and prevents dispersion of the surface soil aggregates, thus maintaining the percolating waters free of suspended material (Lowdermilk, 1930). Unincorporated organic

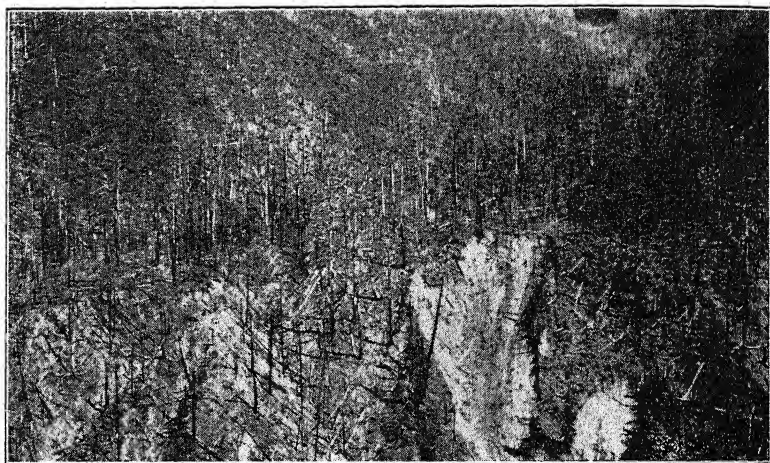


FIG. 47. A portion of the Cispus burn, Columbia National Forest, Washington, showing erosion in the foreground. Photograph by United States Forest Service.

matter and a dense stand of living plants also reduce the rate of overland flow of water, thus permitting more time for infiltration to occur. Infiltration is also favored by the delay of melting of snow in the spring; under forest stands this delay consumes from a few days to 5 weeks. Old root canals and the burrows of soil animals furnish many channels for the passage of water. Under forest vegetation the soil may not freeze as early or to as great a depth as it does in the open. Moreover, forest soils having a large noncapillary pore volume are not impervious to water, even when they are frozen.

2. The surface accumulation of organic matter has a high water-holding capacity, which increases the amount of water that the soil will hold.

3. The root systems of the vegetation hold the soil mechanically, hindering its removal. This situation has been clearly shown by the work of Kramer and Weaver (1936) and Shively and Weaver (1939).

4. Protection against wind is afforded, thus preventing blowing of the soil.

Serious erosion is definitely related to the removal of native vegetation or to disturbances which decrease stand density and damage the soil (Figures 47 and 48). Fires are an important

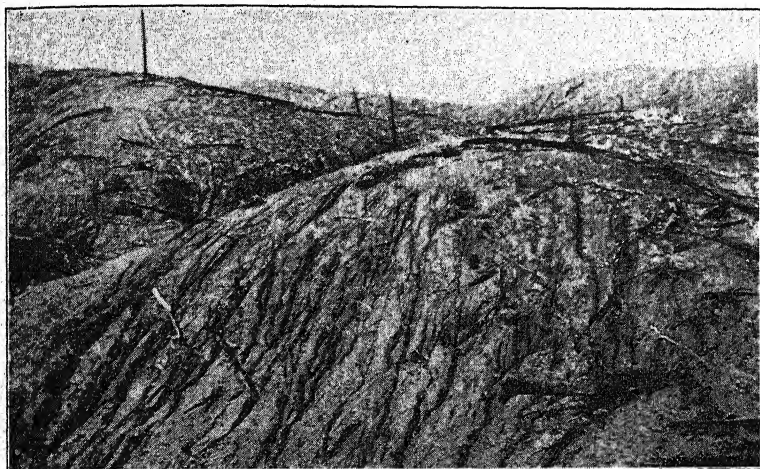


FIG. 48. Severe erosion has occurred after destruction of the forest cover (Squaw Creek, California). Photograph by United States Forest Service.

cause of accelerated erosion in forests (Connaughton, 1935; Munns and Sims, 1936; Anon., 1940; Hendricks and Johnson (1944); Sampson (1944). Kotok (1931) reported that burning of vegetation and litter in brush field areas in southern California resulted in an increase of erosion up to 1000 times that on unburned areas. Munns and Sims stated that during a disastrous flood in Los Angeles County, California, in 1934 from 50,000 to 67,000 cu. yd. of debris was eroded from each square mile of a burned-over area. In San Dimas canyon, which was not burned, only 56 cu. yd. was eroded per square mile.

Excessive grazing in forests or on range lands reduces vegetation density and causes soil changes that lead to increased run-off and erosion. Munns *et al.* (1938) reported that a storm of 4

hours' duration in Wisconsin resulted in the erosion of 745 lb. of soil per acre in a grazed woodlot, but only 17 lb. per acre were removed from an ungrazed woodlot. Investigations by

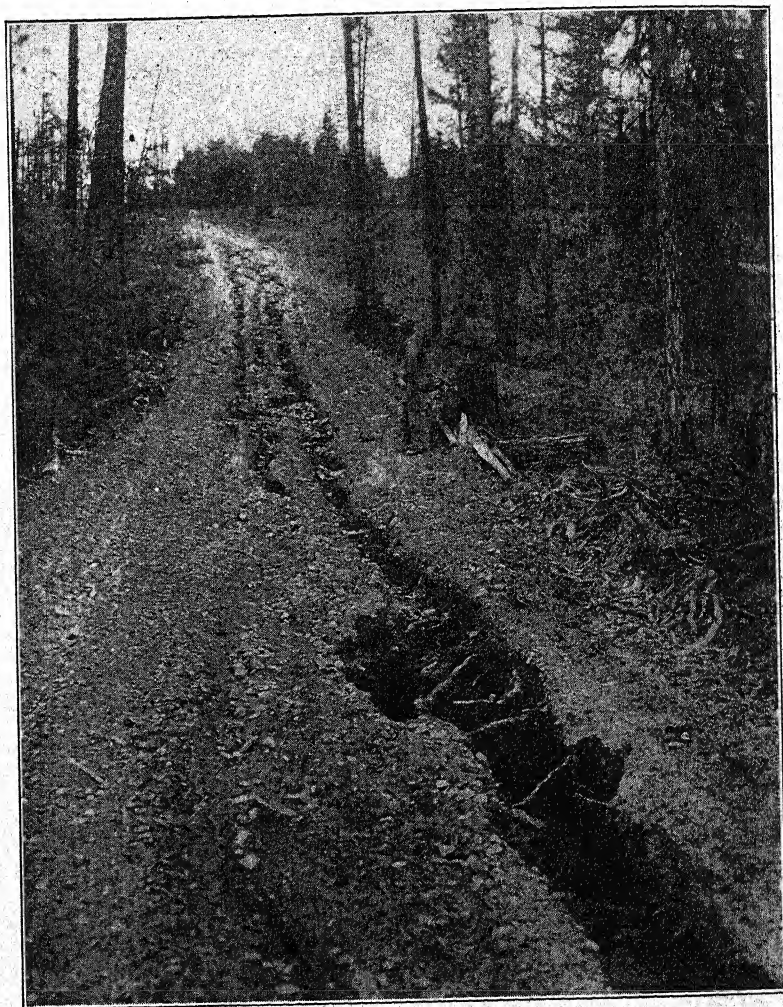


FIG. 49. Gully erosion on a caterpillar logging road in Cabinet National Forest, Montana. Photograph by United States Forest Service.

Renner (1936) on the Boise River watershed in west central Idaho, by Forsling (1931) on the Wasatch Plateau in Utah, by Cooperrider and Hendricks (1937) on the upper Rio Grande

watershed in Colorado and New Mexico, and by Woods (1944) in the intermountain region all point clearly to an intimate relationship between impairment of the range vegetation by overgrazing and accelerated soil erosion. Woodward and Craddock (1945) in an analysis of conditions on range lands in the Sevier Lake basin of south central Utah have indicated a remarkable relationship between the density of plant cover and surface run-off. For example, the recurrence interval of run-off from land with soil derived from limestone and supporting browse range vegetation having a plant density of 5 per cent is estimated to be 1.5 years.¹ With increase in plant density to 15 per cent the recurrence interval of runoff is increased to 50 years. In the Southwest, where erosion is particularly destructive, some investigators have advanced the view that the destructive processes are caused by a change in climate. Cooperrider and Hendricks (1937) examined this theory and concluded that it was not tenable. They, with Rich (1911), Duce (1918), Chapman (1933), and Thornthwaite *et al.* (1942), attribute the accelerated erosion to impairment of the native vegetation, principally through overgrazing.

Logging has, in some places, resulted in accelerated erosion, particularly along skid roads (Figure 49). As a general rule, however, serious erosion does not occur unless fires follow or the area lies bare for a period of several years. The second situation seldom prevails. Even on clearcut areas a protective cover of vegetation usually develops within a year or two after removal of the stand.

THE INFLUENCE OF CULTIVATION AND CROPPING PRACTICES

The most serious soil erosion in America has probably occurred on lands which have been used for agriculture. Native vegetation was destroyed and replaced by crop plants which occupied the land only part of the year. Clean-tilled crops were grown on erosive soils with little consideration of the soil losses which resulted. Joel (1937) reported that 79.3 per cent of all cultivated lands and 89.4 per cent of all idle lands, but only 26.9 per cent

¹ The recurrence interval of run-off is the average period, in years, between storms having intensity-duration patterns that can be expected to produce overland flow from specific land areas. It is based on the anticipated frequency of storms with a rainfall intensity which will exceed the total infiltration capacity of the soil.

of all pasture lands, in the southern short grass plains have been affected by serious erosion. Conrey *et al.* (1937) cited data for widely separated important soil types, which indicated that the annual losses of soil, in tons per acre, varied from 0.003 to 1.300 for areas supporting thick-growing crops and from 13.8 to 60.8 for areas supporting intertilled crops. Enlow (1939) indicated that the average soil loss, in tons per acre annually, amounted to 88.35 when continuous crops of corn were produced on Clinton silt loam near La Crosse, Wisconsin. When a corn, clover, and barley rotation was used, soil losses averaged 25.31 tons.

DAMAGE RESULTING FROM WATER EROSION

The most obvious damage resulting from water erosion is the removal of part or all of the top soil from the eroded areas. It is well known that erosion is frequently selective, removing from the upper soil layers their most valuable constituents. In an investigation in New Jersey Martin (1941) found that eroded material usually contained three to eight times as much organic matter and nitrogen as did the uneroded soil. In addition to loss of organic matter and nutrients erosion usually results in a breakdown of soil aggregates, a reduction in field capacity, a reduction in the rate of infiltration of water, and a decrease in activity of the soil flora and fauna. Croft *et al.* (1943) measured the effects of accelerated erosion of the soils of three range watersheds on the west face of the Wasatch Mountains east of Salt Lake City. On areas where acceleration of erosion was moderate, the surface inch of soil showed an average loss of 43 per cent organic matter and 39 per cent nitrogen and a 24 per cent decrease in moisture equivalent (based on comparisons with soil conditions in areas where normal geologic erosion occurred). On areas where acceleration of erosion was severe the corresponding losses were as follows: organic matter 70 per cent, nitrogen 67 per cent, and moisture equivalent 40 per cent.

Damage frequently results from the deposition of coarse materials, such as sands, gravels, and cobbles, on productive valley lands. Large amounts of hardwood timber in the Mississippi River bottoms died after deposition of silt and sand to depths of from 6 in. to 3 ft. or more.

Another type of damage which is related to accelerated water

erosion is that caused by deposition of silt and sand in stream channels, harbors, reservoirs, and irrigation ditches. According to Renner (1936), it is estimated that 7500 acre-feet of silt became lodged behind the Arrowrock Dam on the Boise River in west central Idaho during the period between 1915, when the dam was completed, and 1927. It is significant that Renner found widespread evidence of accelerated erosion on the Boise River watershed. The damage resulting from silting of reservoirs in other parts of the country has been discussed by Eakin (1936), Barnes *et al.* (1939), Anon. (1940), Garin (1941), and Brown (1943). Investigations conducted by the Soil Conservation Service indicate that the rate of silting can be very greatly reduced through the introduction of soil-conservation practices. Surface run-off is materially greater on eroded lands than on noneroded lands. Consequently it is reasonable to assume that erosion contributes to increased flood heights.

DAMAGE RESULTING FROM WIND EROSION

Damage from wind erosion involves chiefly removal of fertile surface soil, covering of productive soils with material of low fertility, mechanical injury to agricultural crops, burying of forest stands, and injury to human beings and livestock. Figure 50 shows the burial of a forest stand by dune sands. In humid regions wind erosion is localized in areas where the soils are sandy and wind velocities high, but in arid and semiarid regions it is more general. Various aspects of wind-erosion damage have received detailed treatment by King (1894), Free (1911), Joel (1937), Whitfield (1938), and Chepil (1945a, 1945b).

CONTROL OF EROSION

The most effective control of erosion is usually obtained not by application of any one method alone but rather by the integrated use of several methods. The precise techniques appropriate for the control of erosion in different situations vary so greatly that only general principles will be presented. Details of application, particularly as they relate to construction of terraces and dams, may be found in manuals and other publications of the federal and state governments.

As a general principle it is much easier to prevent the initiation of accelerated erosion than to stop it once it has become serious.

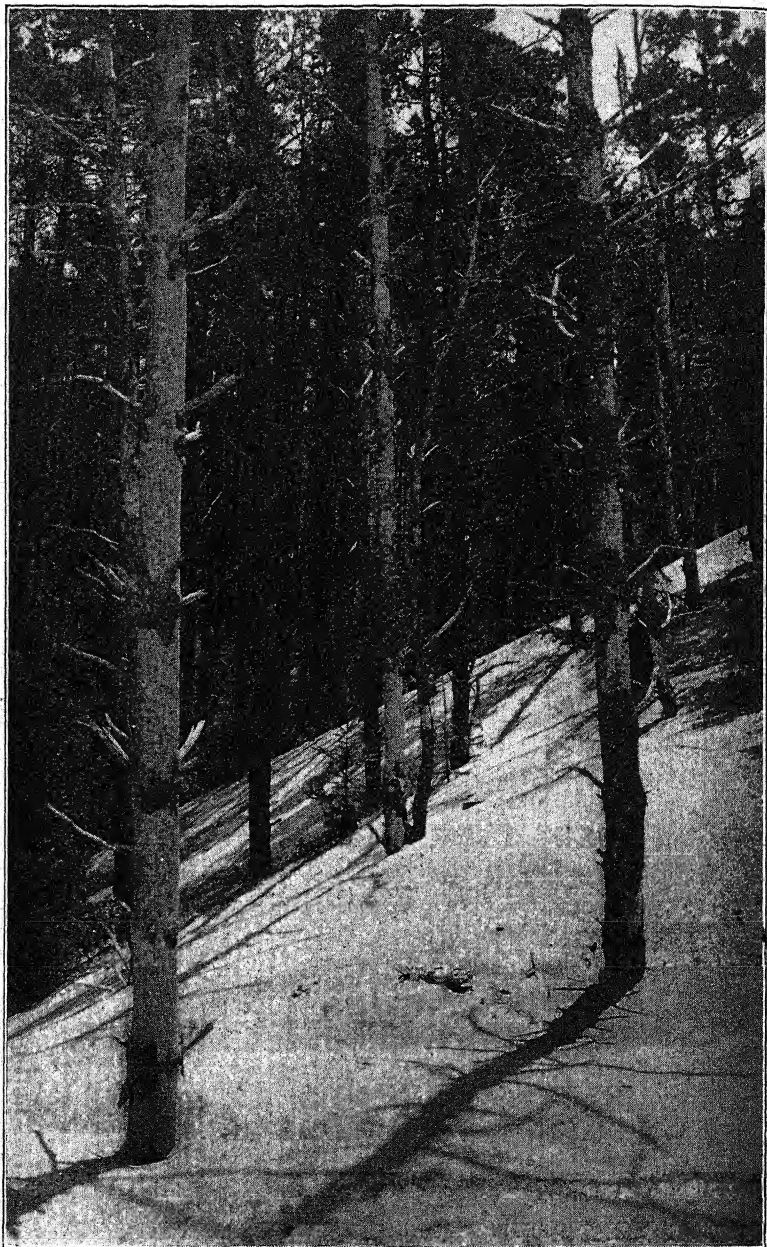


FIG. 50. A forest being slowly buried by dune sand near Siltcoos Lake, Siuslaw National Forest, Oregon. Photograph by United States Forest Service.

Consequently sound land-use practices are needed on areas not now eroding in order that they may continue to be free from this evil. Erosion control should not be thought of solely in terms of arresting erosion after it has started. The control which is most successful is that which prevents inception.

THE USE OF VEGETATION IN EROSION CONTROL

The best protection against erosion is afforded by a vigorous plant cover. Consequently in any integrated system of erosion control vegetation cover is a prime consideration. Mechanical devices or techniques are extremely valuable when used to supplement measures employing vegetation; however, such mechanical devices should not be regarded as a substitute for adequate vegetation.

Forest vegetation is one of the most efficient types of plant cover in erosion control. Therefore, in regions where climatic conditions are favorable, reforestation should play an important role in the management of lands which have been severely eroded and which cannot be safely continued in cultivated crops. It is estimated that some 50 million acres in America have been permanently released from cultivation and that an additional 30 million acres will be abandoned in the near future. Of these lands about 21 million acres will have to be reforested for control of erosion (Munns *et al.*, 1938). Trees are particularly valuable on areas where erosion is extreme and where permanent maximum protection is essential. The trees best adapted for planting on severely eroded sites usually are native species which are intolerant and which have relatively low moisture and nutrient requirements.

Wind erosion may be effectively controlled in humid climates by reforestation of the shifting sands (Altpeter, 1941; Lehotsky, 1941; McLaughlin and Brown, 1942). First it is usually necessary to stabilize the soil material by planting beach grasses or by providing mechanical barriers, such as brush or pickets. Establishment of a forest cover by planting is the second step in effecting permanent control. Hardy conifers native to the region are generally used. Windbreaks of trees have been found helpful in the control of wind erosion on the Great Plains of the West (Stoeckeler, 1938).

In established forest stands erosion can best be prevented

through the application of sound forest practices. This program usually involves cultural operations and harvest cuttings but at the same time requires maintenance of adequate cover to protect the soil. Injurious influences, such as fire and excessive grazing, must be rigorously excluded. Stands on areas where run-off is appreciable and the danger of water or wind erosion is great, require more careful management than those in less critical situations.

In large sections of this country climatic conditions make it impractical to use trees in erosion control on lands released from cultivation. In such situations revegetation with grasses and shrubby vegetation is indicated. Even in humid regions where trees can be grown it is commonly desirable to restock eroding lands with grasses or shrubs in order to meet the needs of the landowner and to create favorable environments for wildlife (Stevens, 1937). Where a possibility of choice among trees, shrubs, and grasses exists, the final decision concerning the type or types of vegetation to be used should be based on a careful evaluation of the local situation. Many species of grasses and other plants which form dense cover are available for use in erosion control (Enlow and Musgrave, 1938). Revegetation is greatly favored by seeding and fertilization.

In natural grasslands the requirements for erosion control depend on the condition of the range. If the native plant cover is not depleted as a result of misuse, proper regulation of grazing is usually sufficient to prevent accelerated erosion. On the other hand, if deterioration has already reached an advanced stage, mere restriction of grazing is not enough. Additional measures are then necessary to hasten revegetation and reduce run-off. Because of the cost, artificial restoration of vegetation on all damaged grazing lands is impossible. Consequently restorative measures, in addition to the restriction of grazing, should first be applied to two classes of land: (1) critical areas where community welfare demands immediate control of soil erosion and run-off, and (2) areas of valuable spring and fall range where soil conditions are favorable for forage-plant growth. Reseeding of these areas with species of proven adaptability appears to be both feasible and essential (Cooperrider and Hendricks, 1937; Stewart *et al.*, 1939; Flory and Marshall, 1942; Pearse and Hull, 1943). Supplementary measures, such as contour

furrows, absorption terraces, and check dams in gullies, contribute materially to the establishment of vegetation. In especially critical areas fertilization, seed-bed preparation, and even transplanting may be justified.

On cultivated lands which are subject to erosion, strip-cropping practices are highly recommended (Kell, 1938). *Contour strip cropping* refers to the production of agricultural crops in long, relatively narrow strips of variable width running across the slope, approximately on the contours. Strips of densely growing, soil-protecting crops are alternated with strips of inter-tilled crops. *Field strip cropping* is a modification of contour strip cropping. The parallel strips are generally uniform in width and do not follow the contours, although they are laid out crosswise of the general slope. *Wind strip cropping* consists of the production of agricultural crops in long, relatively narrow, straight, parallel strips placed crosswise of the direction of prevailing winds. Strip cropping on sloping land increases infiltration and reduces the velocity of run-off water, and the strips of densely growing vegetation tend to hold the soil washed from the clean-tilled strips. It has been demonstrated repeatedly that amount of run-off and erosion is much less from strip-cropped land than from lands where this technique is not used.

Continuous production of a single crop, particularly one requiring clean cultivation, such as cotton, corn, or tobacco, is known to result in serious soil-erosion losses. These losses can be minimized by crop rotation, which involves the growing of different crops in recurring succession on the same land. In addition to lessening erosion, rotation of crops has other advantages over the one-crop system; for example, it aids in controlling weeds, insects, and plant diseases and favors the maintenance of soil organic matter and nitrogen. The crop rotation commonly recommended is (1) a tilled crop, (2) a small grain, and (3) a crop of grass or legumes. Detailed information on appropriate rotations for different sections of the country are presented by Leighty (1938) and Enlow (1939).

In addition to the foregoing measures for lessening erosion on cultivated lands it is advisable to make the greatest possible use of crops having high protective value. Crops of this kind include the various grasses and legumes. Cover crops, which cover and protect the soil, especially in winter, and green manure

crops, which add organic matter to the soil, are both effective in erosion control.

Wind erosion in the Great Plains may be materially lessened by changes in land use. Overgrazing of the pasture lands must be avoided, and crop residues should be left on the land as a protective cover. Tillage methods, such as listing, which roughen the soil surface and increase absorption of precipitation aid greatly in preventing wind erosion. Chilcott (1937) has pointed out that the prevention and control of soil blowing are dependent on maintenance of a protective cover of crops, crop residues, or clods on the land surface.

THE USE OF VEGETATION IN GULLY CONTROL

Control of gullies requires the use of vegetation. Jepson (1939) has pointed out that the objective of gully control is, in fact, the development of an erosion-resistant cover of vegetation which will stabilize the soil and also produce a useful crop. The principal value of most mechanical structures in erosion control lies in their facilitation of revegetation of the eroded lands.

Occasionally natural revegetation of gullies may be sufficiently rapid to insure reasonably prompt development of a protective cover. More commonly, however, natural revegetation is so slow that planting or reseeding is necessary. Species for use in gully control must be selected with great care if costly failures are to be avoided (Ligon, 1940). Experience indicates that native species are more likely to succeed than exotic species obtained from regions having different climatic conditions. Exotics should be used only when they are known to be adapted to the environment. Not all native species are suitable for planting on eroded lands... Those characteristic of pioneer communities are usually better adapted to the severe conditions found on gullied areas than are species characteristic of climax communities. Attempts to establish stands of climax forest species on eroded lands in the Tennessee Valley have almost always been unsatisfactory. For lists of species advised for gully control in various parts of America the reports of Kraebel and Pillsbury (1933), Stevens (1937), Enlow and Musgrave (1938), and Jepson (1939) should be consulted.

Before planting is attempted, mechanical measures are usually

taken to divert the run-off water before it enters the gully; if this diversion is not feasible, dams of various types are constructed in the gully to prevent further cutting. In addition to these measures, which are described in a subsequent section, it is sometimes necessary to prepare for planting by breaking down the gully banks (Meginnis, 1933; Jepson, 1939). This type of preparation is expensive and should be employed only when the banks are too steep to plant. The grading or sloping may be accomplished by plowing, by dynamiting, or by hand labor in the fall or early winter. A further step to be taken is the exclusion of livestock from the area in order to avoid damage to the developing vegetation.

If trees are to be planted, the stock should be strong and healthy; transplants are commonly superior to seedlings. Spacing for trees varies with local conditions and the species used but is seldom more than about 6 by 6 ft. Shrubs are planted more closely, from about 3 by 3 to 4 by 4 ft. In small gullies it is sometimes possible to use shrubs in forming checks across the gully bottom. Water velocities are thereby reduced and silt is collected, making the development of other vegetation easier. The shrubs are spaced about 3-6 in. apart in shallow trenches extending completely across the gully floor. A row of stakes driven in the ground about 1 ft. down the channel from the shrub check affords some protection and causes the deposition of silt around the plants. Vines such as kudzu (*Pueraria thunbergiana*) and the Japanese honeysuckle (*Lonicera japonica*) have been used successfully in gully control, but they are usually not planted with trees or shrubs.

Grasses and legumes are effective in controlling erosion in small gullies. However, some seed-bed preparation and fertilization are usually required. After seeding, a mulch of straw, brush, or other material contributes greatly to successful establishment of the plants (Franklin, 1939). Hendricks (1938) reported satisfactory revegetation of small gullies in the Southwest by means of burlap sacks containing seeded soil. Seeds of native grasses were mixed with the soil next to one side of the sack, which was then placed with the seeded side up on a spot where the soil of the gully bottom had been loosened. Sod may be transplanted in a gully when the volume of run-off or the necessity of immediate protection makes seeding impractical.

This method of establishing a protective cover is not extensively used, however, because of its high cost.

MECHANICAL MEASURES IN EROSION CONTROL

In many situations mechanical measures are necessary to supplement vegetation in erosion control and to aid the establishment of a plant cover. By means of relatively simple techniques it is possible to increase materially absorption of water by the soil, to reduce the velocity of run-off, and to provide for the movement of large volumes of water through protected channels. Frequently it is necessary to estimate the probable rate and amount of run-off from a given drainage area in order to ascertain the type and size of structures to be used. A table for estimating probable run-off from various types of watershed is presented by Jepson (1939: 12).

INTERCEPTION AND RETENTION OF WATER

Contour trenches or furrows afford a simple means of retaining and distributing water on slopes. They have found considerable use on pasture and range lands (Bailey and Croft, 1937; Nichols and Chambers, 1938; Jepson, 1939). The trenches or furrows vary in size and spacing, depending on the slope, the run-off, and the amount of water to be impounded. Sometimes only shallow furrows are constructed, and water in excess of their storage capacity is allowed to flow down the slope. If local conditions require that all but the highest anticipated run-off be impounded, it is necessary to provide trenches of adequate size and number to supply the desired storage capacity. Equalizers or small earth dams are thrown up in the trenches at intervals of 20-100 ft. to divide the trench into a series of small storage units. The equalizers are so constructed that they are somewhat lower than the top of the trench; this arrangement permits water from filled trench segments to flow laterally into unfilled units. Furthermore, should a break in a trench wall occur, the equalizers prevent loss of all the impounded water. Earth dams having a height equal to that of the trench walls are also placed at gullies and at points along the trench where changes in grade occur. The purpose of these dams is to cause each trench section to function independently.

Basin listing, which is accomplished with a machine, consists

of creating small earth dams across the lister furrows at intervals of 15-25 ft. This technique, which is commonly applied to fallow agricultural land in dry regions, provides for the retention and absorption of practically all precipitation. Wind erosion is greatly lessened on lands which have been basin-listed or strip-listed.

Absorption terraces function in erosion control by reducing run-off. They are constructed so as to spread the collected run-off over as wide an area as possible. Terraces of this type are most useful in the central Great Plains on areas where slopes are gentle and the soils absorptive (Hamilton, 1938).

Bench terraces are used to a limited extent in America to permit crop production on steep slopes. They consist of a series of level or nearly level benches separated by nearly vertical risers. Cultivation of land on slopes so steep as to require bench terraces can be justified only in exceptional cases.

A series of earth fills are occasionally used to impound water in small gullies on cultivated lands having absorptive soils. The fills are extended above the top of the gully, and provision is made for the overflow to be diverted from alternate sides. In time the storage capacity of the gully is reduced as a result of deposition, and overflow then increases. Consequently, if erosion is to be permanently arrested, it is necessary to develop a cover of vegetation which will protect the soil.

Various other measures may be applied on cultivated lands to increase retention of precipitation. Subsoiling consists of loosening the subsoil, usually by means of special machinery, to a depth of 1-2 ft. in order to render it more pervious to water. Tillage should be along the contour, not up and down slopes. The furrows resulting from contour tillage retain water until it can be absorbed by the soil, but furrows running up and down slopes favor increased run-off and erosion.

DIVERSION OF RUN-OFF WATER

On steep or unusually long slopes the amount of run-off may be so large that erosion losses become important. The diversion of run-off water by means of drainage terraces or diversion ditches should then be considered. However, it is always necessary to provide for safe disposal of the water concentrated in the drainage or diversion channels. Failure to provide adequately

for disposal of the run-off is certain to result in new damage. Adjacent forest lands or pastures, if in good condition, afford suitable places to empty the diverted water. Care should be taken, however, that the velocity of flow does not exceed safe limits. It may be necessary to spread the flow over a substantial area in order to increase infiltration and avoid erosion. When areas well protected by forest or grass are not available for disposal of the diverted water, natural or artificial channels must be utilized. Artificial channels are classed as either low-velocity or high-velocity. Low-velocity channels are relatively wide, shallow channels which are protected by a dense cover of perennial vegetation, such as grass, or sometimes low shrubs. High-velocity channels are used on steep grades where low-velocity channels are impractical. This type of channel is relatively narrow and is constructed of rocks, concrete, or other materials which prevent erosion of the bed and sides.

Drainage-type terraces are structures whose primary purpose is to conduct run-off water from fields at nonerosive velocities (Hamilton, 1938). The terrace consists of a low ridge with a wide, relatively shallow channel on the upper side. The gradient of the channel should be low, so that deposition, and not erosion, occurs. Tillage operations are carried out on the slopes of both the terrace ridge and the channel. Terraces of the drainage type are most widely used in regions where soils have rather low permeability and where precipitation is abundant, for example, in the southeastern and middle Atlantic States.

Diversion ditches may also be used to conduct run-off water from cropland or from other lands which are being eroded. The gradient of the diversion ditches, like that of the drainage terraces, must be low enough to prevent erosion, and a suitable outlet must be provided. It is considered good practice to locate each channel immediately below an area supporting a good cover of vegetation, such as forest or grass. This practice reduces the amount of silt which washes in and consequently lowers maintenance costs.

CONTROL OF GULLIES

It is easier to prevent gullies from forming than to control them once they have developed. In general, gullies do not form on properly managed lands. The adoption of good practices in

forest management, range management, and agriculture is the most certain way to avoid the development of serious erosion of the gully type. The forester, grazier, and agriculturist should be well informed as to the condition of the lands under his supervision and, at the first signs of accelerated erosion, should take appropriate steps to stop the damage before it attains serious proportions. Only in this way can expensive erosion control be avoided.

Gully control is greatly facilitated if measures are taken to increase retention of water on the area and to provide for diversion of run-off away from the gully head. Practices by which these purposes may be achieved have already been considered. Sometimes diversion of run-off may be the only mechanical measure required for control. Vegetation can then be established to make the control permanent.

If it is not feasible to keep run-off water out of a gully, it becomes necessary to install check dams or other structures to halt further cutting until vegetation can become established. Another purpose served by check dams is the accumulation of soil material in which trees or other vegetation can be planted.

Temporary Dams. Small temporary check dams are frequently needed in gullies to aid the establishment of vegetation. The life of these structures should be from 3-8 years. They may be constructed from heavy-gage woven wire, brush, poles, loose rock, planks, or other materials which are readily available (Nichols and Chambers, 1938; Meginnis, 1938; Jepson, 1939). Although check dams are temporary, they must be carefully designed and constructed if they are to function effectively during the years when vegetation is being established (Figures 51 and 52).

Low dams with an overfall height of about 1.0-1.5 ft. are preferable to higher structures. Good anchorage to the gully bottom and sides must be secured in order to prevent washing underneath or at the sides. Sufficient spillway capacity must be provided to accommodate the maximum flow which is anticipated during the life of the dam. In order to avoid undue concentration of water at the overfall, spillways should be made as broad as is practical. Undercutting at the overfall may be prevented by providing an apron built of rocks, brush, logs, planks, or sod.

Meginnis (1939) found soil-collecting trenches a promising sub-

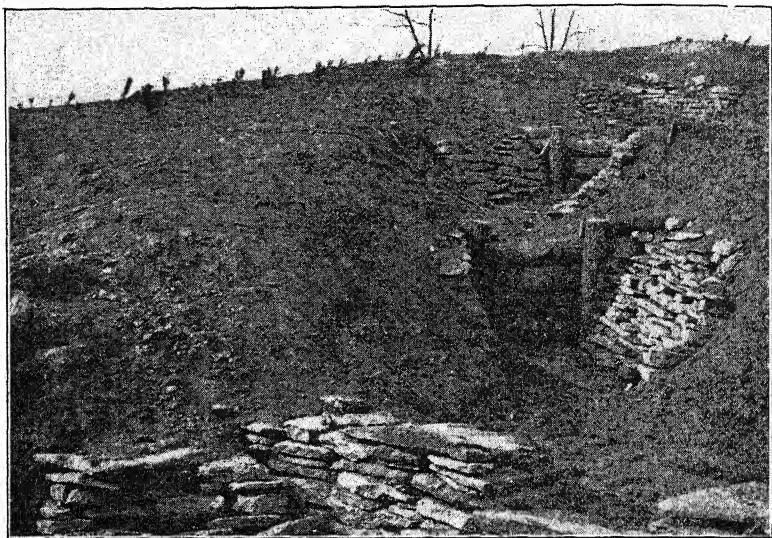


FIG. 51. Check dams built of logs and rocks are used to control erosion in this gully in Clermont County, Ohio. Photograph by United States Forest Service.

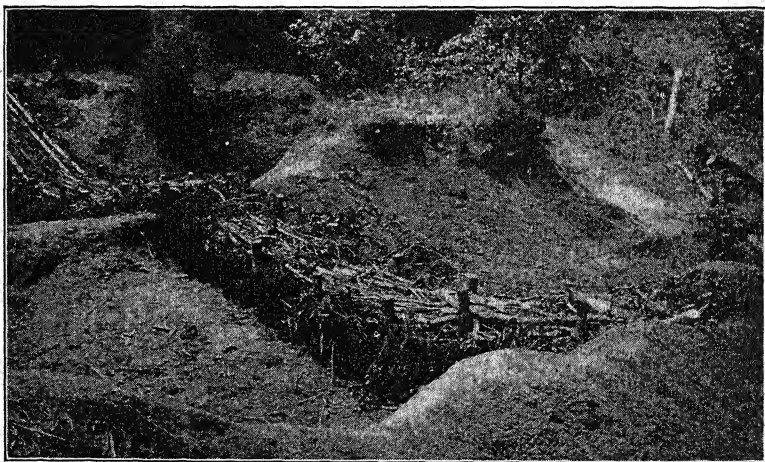


FIG. 52. The check dams shown in this view were built of brush, stakes, and wire (Carroll County, Tennessee). Photograph by United States Forest Service.

stitute for check dams in reforesting gullies in Mississippi. Trenches about 18 in. wide and 18 in. deep were constructed at right angles to the line of flow with the excavated material piled down slope. Two years after the trenches had been filled by washing in of sediment, they were planted with loblolly pine and black locust. The accumulated soil favored good survival and rapid growth of the trees.

Permanent Dams. Large gullies having a vertical drop at the head frequently require permanent structures to halt erosion. Such dams are usually constructed of concrete, masonry, or earth. Permanent dams represent a substantial investment, and their design and construction require considerable technical knowledge. Consequently it is desirable that the service of a competent engineer be obtained before work is initiated.

Two types of dams are commonly used, namely, notch spillway and drop inlet dams. Notch spillway dams are usually constructed of concrete or masonry, and drop inlet dams of earth (Kessler, 1934; Nichols and Chambers, 1938; Jepson, 1939). Permanent dams are located at or near the gully head. The spillway crest must be high enough to pond water across the gully overfall.

Flumes. Flumes or chutes are used to conduct run-off water down gully-head overfalls or other steep pitches. They are constructed of concrete, masonry, wood, metal, or sod. Walls or dikes are built at the flume entrance to guide the flow to the inlet, and protective devices are installed at the outlet to prevent damage by the discharging water. Outlet protection is obtained by means of an apron or other structure which serves to reduce the kinetic energy of the water before it is discharged on the gully floor.

CONTROL OF EROSION ALONG ROADS

Drainage water along roads is a cause of much damage unless proper provisions are made for its disposal. Gullies may originate at either the inlet or the outfall of a culvert and destroy adjacent valuable land. To prevent erosion of this kind drop-inlet culverts with protected outfalls may be used. Flumes or chutes are useful in conducting run-off down steep cut or fill slopes.

Kraebel (1936) has pointed out the serious erosion conditions

which result from construction of modern high-standard roads in mountain regions. He indicates that the erosion hazard of new roads can be materially lowered by more rational alignment, greater use of retaining walls and cribbing on cuts and fills, more care in disposing of excess material, occasional use of tunnels instead of extremely deep cuts and of bridges instead of large fills, and improved drainage practices. A protective cover of vegetation should be developed for permanent control on cut and fill slopes. Before seeding or planting, the slopes may have to be smoothed and anchored with wattles and stakes. The stabilization of roadside slopes by vegetation has been discussed by Kraebel (1936), Hendricks and Grabe (1939), Hursh (1939), and Johnson and Brown (1941).

CONTROL OF EROSION ALONG SMALL STREAMS

Cutting of banks on small streams can usually be controlled by simple measures involving the use of mechanical structures and vegetation. Jetties or retards built out into the stream from the eroding bank check the velocity of flow and cause deposition of eroded material. These structures are pervious to water and are built of rocks, brush, or logs. In streams having a high velocity, adequate provision must be made for anchorage. As a general rule, the length of the individual jetties should not be more than about one-quarter or one-third of the stream width at flood stage. They may be spaced at a distance about equal to the stream width. After bank-cutting has ceased, willows, cottonwoods, alders, or other vegetation should be planted for permanent protection.

Revetment of stream banks with loose rock or brush matting may be cheaper than construction of jetties. In deciding on the control measure to be taken, the matter of cost should always be considered. If the revetment is temporary, vegetation should be established in order to make the control permanent. Control of erosion on the streambanks of the Winooski River in Vermont has been accomplished through the use of rock riprap on the lower slope of the banks to a point approximately 5 ft. vertically above normal summer low-water level and the establishment of woody vegetation on the upper part of the slope (Altpeter, 1944).

Stabilization of eroding stream banks is hastened by excluding domestic livestock and avoiding cultivation of the land immediately adjacent.

FOREST SOIL DETERIORATION

It is generally accepted that soil conditions are not static but are subject to change. Changes in soil properties are of interest to the silviculturist, because they may either favor or hinder the production of timber crops. In the preceding pages various aspects of soil change have been discussed, but there remains the question of soil deterioration resulting from pure stands and clearcutting. These influences are frequently held to be damaging to forest-soil productivity.

SOIL DETERIORATION RESULTING FROM PURE STANDS

The idea that pure forest stands, particularly when grown as successive crops on the same land, necessarily lead to soil deterioration is rather widespread among foresters in this country. It is a concept which frequently arises during discussions of the relative merits of mixed, as compared to pure, stands. Many American foresters evidently believe that the unfortunate experience of the Germans with certain pure spruce stands in Saxony justifies the broad generalization that pure stands of any conifer on any site seriously damage the soil and consequently represent bad silviculture. The circumstances other than purity of the crop which contributed to the difficulties in Saxony are frequently overlooked.

The rocks from which the forest soils of Saxony developed are, in general, poor in calcium. Krauss (1928: 343) stated that the low content of calcium in the soils may be regarded as a primary cause of the tendency for organic debris to decompose slowly and produce acid humus layers. Climatic conditions in Saxony are not particularly favorable for the development of spruce, especially in pure stands at low altitudes. Wiedemann (1925: 35) has pointed out that the mean annual precipitation, the number of drought months, and other measures of moisture relations

all point to critical conditions in the lower-lying districts, as follows:

	Altitudinal Belt, feet above sea level				
	1970-2620		1150-1970	660-1150	< 660
	Absolute Value	Per-centage	Per-centage	Per-centage	Per-centage
Mean annual precipitation, inches	37.7	100	82	77	67
Number of drought months (< 1.6 in. rain), May to September, 1911-20	4.5	100	198	249	364

Krauss (1928: 348) stated that Saxony has fewer heavy rains than almost any other region in central Europe. When it is recalled that dense spruce stands intercept from one-quarter to one-third of the precipitation which falls, that spruce tends to be shallow-rooted and dependent on summer rainfall, and that its water requirements are relatively high, the decline in growth of pure stands in lowland situations during drought periods, for example, 1911-21, is not surprising. Even in the upper altitudinal zones in the Erzgebirge, where spruce is adapted to the site and where moisture is sufficient, the species does not reach its optimum development because of low temperatures.

Many additional factors have had a bearing on the present unsatisfactory conditions noted in some of the Saxon pure spruce stands. Human influences, such as ruinous exploitation of the forest for charcoal wood and litter removal, had far-reaching effects in forests of the Erzgebirge much earlier than in other less highly industrialized regions. Under the influence of the theory of highest net revenue and the highly developed Saxon pulpwood industry, pure spruce was planted more and more. It was even established in the lowland hardwood regions and on pine sites where neither climatic nor soil conditions favored its natural occurrence. Furthermore, little attention was paid to the source of seed and planting stock, with the result that unsuitable strains were undoubtedly used in many plantings. Faulty management practices in the stands, undue exposure of the site, removal of stumps from clearcut areas, and industrial smoke are all known or believed to have exerted unfavorable influences. Putscher (1928: 391) observed that the widely practiced stump removal on heavy-textured soils in Saxony resulted in greater

compactness of the soil, and that this practice alone must have led to lower production in the second generation stands. As both Wiedemann (1928a: 397) and Krauss (1928: 350) have emphasized, the effects of practices and conditions of the past should not be underestimated when judging present-day forests.

The results of excessive use of spruce in the middle elevations, where pure stands of the species were generally not adapted to the site, and particularly its use in the lower elevations led to deterioration of both stands and soil. Water relations were unfavorably affected as a result of the interception of light summer rains by the tree crowns and the mor humus layers which developed over the mineral soil. Damage from drought and disease increased. In 1919 Wiedemann began a thorough investigation of the reduction in growth and published the results in a series of reports (1923, 1924, 1925, 1928a). His major conclusions were that the falling off of growth was chiefly dependent on unfavorable climatic and soil conditions. In the report published in 1923 it was stated that, although a serious reduction in growth occurred in 70 per cent of the lowland districts, only 16 per cent of the districts in the true spruce zone (above 1800 ft.) of the Erzgebirge were affected. In the state forest considered as a whole, only 7 per cent of the area was occupied by very "sick" spruce stands. In 1925 Wiedemann published a second edition of his 1923 report and took occasion (p. 152) to indicate that the serious deterioration of the mineral soil had been over-emphasized in the earlier publication. In his own words, the deeper he went into the question of soil deterioration, the less justification he found for extreme pessimism. Krauss (1928: 351), one of the foremost authorities on forest soils in Germany, stated that it was an exaggeration of the facts to say that the centuries of spruce culture had resulted in a clearly proven general deterioration of the forest soils in the Saxon state forests.

In 1928 at a meeting of the Society of German Foresters Wiedemann recognized that, as a reaction to the earlier policy of growing spruce in pure, even-aged stands which were clearcut and artificially reproduced, there had developed the view that on every site every pure spruce stand — even every pure coniferous stand — was a promptly acting "poison" for the soil. It was pointed out that this was a greatly exaggerated generalization; on rich soils in Saxony many generations of pure spruce were

known to have followed one another with no appreciable injury to the soil. Wiedemann stated that he had always regretted that the basic data incorporated in his 1923 report had been overlooked in later discussions of the problem and deplored the much too gloomy impression which outsiders had developed of the results of spruce management in Saxony.

Other Europeans who have recognized the fallacy of the generalization that all pure stands are injurious include Bühler (1922: 198), Erdmann (1924: 197), Dengler (1935: 284-285), and Ružička (1938). Dengler observed that, if soil deterioration automatically resulted from pure spruce stands, the boundaries of the spruce region should contract during the course of time. Such contraction, however, has not occurred.

The principal lesson to be learned by Americans from the experience with pure stands of spruce in Saxony and other parts of Europe is that it is dangerous to grow a species outside its natural range on sites to which it is unsuited. Even within the limits of its geographical range a species having the peculiar characteristics of Norway spruce, such as dense crown and shallow root habits, should not repeatedly be grown in pure stands on sites with soil or climatic conditions that render such management unsafe.

In regions of podzol, brown podzolic, and gray-brown podzolic soils a perceptible increase in the rate of podzolization is to be expected when species whose litter is relatively high in bases are replaced by dense, pure stands of species whose litter is low in bases. Various European investigators have reported that podzolization tends to increase when pure Norway spruce stands are established on brown forest soils, and in America Fisher (1928) and Griffith *et al.* (1930) have indicated that a similar change occurs under pure eastern white pine on brown podzolic soils in Massachusetts. Fisher (1928) stated that a soil profile under an 80-year-old eastern white pine stand showed a "strongly podzolized" condition, whereas a profile under an adjacent 18-year-old hardwood forest showed a mull type of humus layer. The view was expressed that the transformation was completed 12-15 years after clearcutting of the white pine. The work of Griffith *et al.* (1930), also carried out in the same region, failed to confirm the existence of the earlier mentioned "strongly podzolized" condition under pine. Until the pine stands were about

40 years old, practically no leached layer could be recognized, and then only 17 per cent of the profiles showed even a trace (< 0.1 in.). In 70-year-old stands no leached layers were found, and in 80-year-old stands 4 of the 10 profiles examined showed no leached layer, 3 had a layer 0.05–0.1 in. thick, 1 a layer 0.1–0.2 in. thick, and 2 a layer 0.4–0.6 in. thick. Of a total of 192 profiles measured in pine stands of all ages only 3, or 1.5 per cent, had a leached layer more than 0.2 in. thick; 77.0 per cent of the profiles showed no leached layer whatever. The view that pure eastern white pine has caused appreciable soil deterioration in Massachusetts is open to serious question. Even less tenable is the generalization that eastern white pine or any other conifer, when grown in pure stands, must invariably or even generally result in soil deterioration.

In many parts of this country *naturally occurring* pure forests (80 per cent or more of the dominants of one species) are to be found. There are no grounds for believing that the soil under these stands has been or will be deteriorated as a result of purity of the crop alone. Artificially established pure stands of conifers are being developed over substantial areas in the eastern United States. Where the species are adapted to the site and are given proper silvicultural care, there seems to be no reason to fear soil deterioration as a result of producing one pure crop. In the majority of stands a considerable number of native trees, shrubs, and herbs are either present on the land at the time of planting or seed in as soon as thinnings or other cuttings are made. Volunteer trees which are deep rooted or which produce litter high in bases and nitrogen should be retained, providing that they do not seriously compete with the main crop. The natural processes of soil leaching cannot be prevented but may be partially counteracted by a small proportion (up to 20 per cent) of these "soil-improving" species in the stand. It is highly probable that many of the essentially pure conifer plantations that have been established in the eastern United States will be followed in the second rotation by either a forest of native hardwoods, naturally seeded, or by conifers in varying mixture with native hardwoods.

Even if none of the native hardwood species develop in conifer plantations during the first rotation—a situation that rarely exists—native herbaceous and shrubby species will seed in and

contribute to the maintenance of soil fertility. The litter of these herbs and shrubs is, in general, relatively rich in nutrient materials and decomposes readily, having much the same effect on the soil as do leaves of hardwood trees. A conifer stand with a rich lesser vegetation may, from a biological point of view, be regarded as a mixed stand.

A slight increase in the rate of podzolization, as has been indicated, may be expected as a result of replacing tree species whose litter is high in calcium and other metallic cations by species whose litter is low in these constituents. The amount of soil change that occurs during one rotation is so small, however, that it need not cause concern. Foresters should recognize that changes in forest composition commonly call forth changes in soil conditions. The silviculturist who develops a pure stand of white pine, spruce, or some other species on land which previously supported a mixed stand should anticipate a change, not necessarily deterioration, in soil properties as a natural accompaniment of such a stand. Although it is unlikely that the forester can by practical means greatly improve soil conditions through regulation of stand composition, it is possible and practical for him to adopt measures that will favor maintenance or even a slight improvement of fertility. These measures, as they relate to stand composition, consist of favoring a representation of "soil-improving" tree species in the stand whenever feasible. Among species in this category the following may be mentioned: *Carya cordiformis*, *C. ovata*, *Celtis occidentalis*, *Cercis canadensis*, *Cornus florida*, *Fraxinus americana*, *Juniperus virginiana*, *Liriodendron tulipifera*, *Liquidambar styraciflua*, *Populus tremuloides*, *Prunus serotina*, *Sassafras albidum*, *Thuja occidentalis*, *Tilia americana*, and *Ulmus americana*.

SOIL DETERIORATION RESULTING FROM CLEARCUTTING

The effects of clearcutting on soil cannot be generalized for all situations. In some instances an unfavorable, and in others a favorable, influence is observed. Foresters should avoid both the unwarranted generalization that clearcutting is usually harmful and the equally untenable view that it is usually beneficial (Wittich, 1930; Wiedemann, 1934; Dengler, 1935: 487).

The length of time that a clearcut area lies exposed has an important bearing on soil conditions. Usually a cover of herbaceous, shrubby, or tree vegetation develops promptly and affords adequate protection against excessively high surface temperatures, deflocculation of the soil aggregates, and erosion. On the other hand, if an adequate cover of vegetation fails to develop for several years after removal of the stand, it is probable that deterioration of the soil will occur. The changes most commonly observed are accelerated erosion and increased compactness resulting from destruction of the soil aggregates and plugging of the noncapillary pores. Fine-textured soils are far more likely to become compacted than are coarse-textured soils. If it is apparent that natural revegetation is too slow to supply the necessary protection to the clearcut area, artificial reforestation should be employed.

As early as 1887 Müller recognized that soils having a mor type of humus layer are frequently benefited by complete removal of the stand. Decomposition of the unincorporated organic matter is hastened, acidity decreases, nitrification is favored, and the soil fauna becomes more active. Wittich (1930) found that the pore volume of sandy soils in northeastern Germany was increased as a result of clearcutting. There is no evidence that clearcutting results in any improvement of soils having a mull type of humus layer. It is, however, rather generally agreed that deterioration occurs if they are subjected to undue exposure.

Swamping has frequently been observed to follow clearcutting of stands on impermeable soils with a very high water-table. The rise of the water-table under these circumstances is believed to result from the elimination of transpiration. This subject has already been considered in Chapter 9.

In recent years there has been a tendency to condemn clearcutting as unconditionally bad practice. Wiedemann (1928b: 682) and many other foresters have deplored this extreme view, which in Europe was carried to the point of characterizing clearcutting as "murder of the forest organism." It should be evident that it is incorrect to regard clearcutting as having either generally unfavorable or generally favorable effects on the soil.

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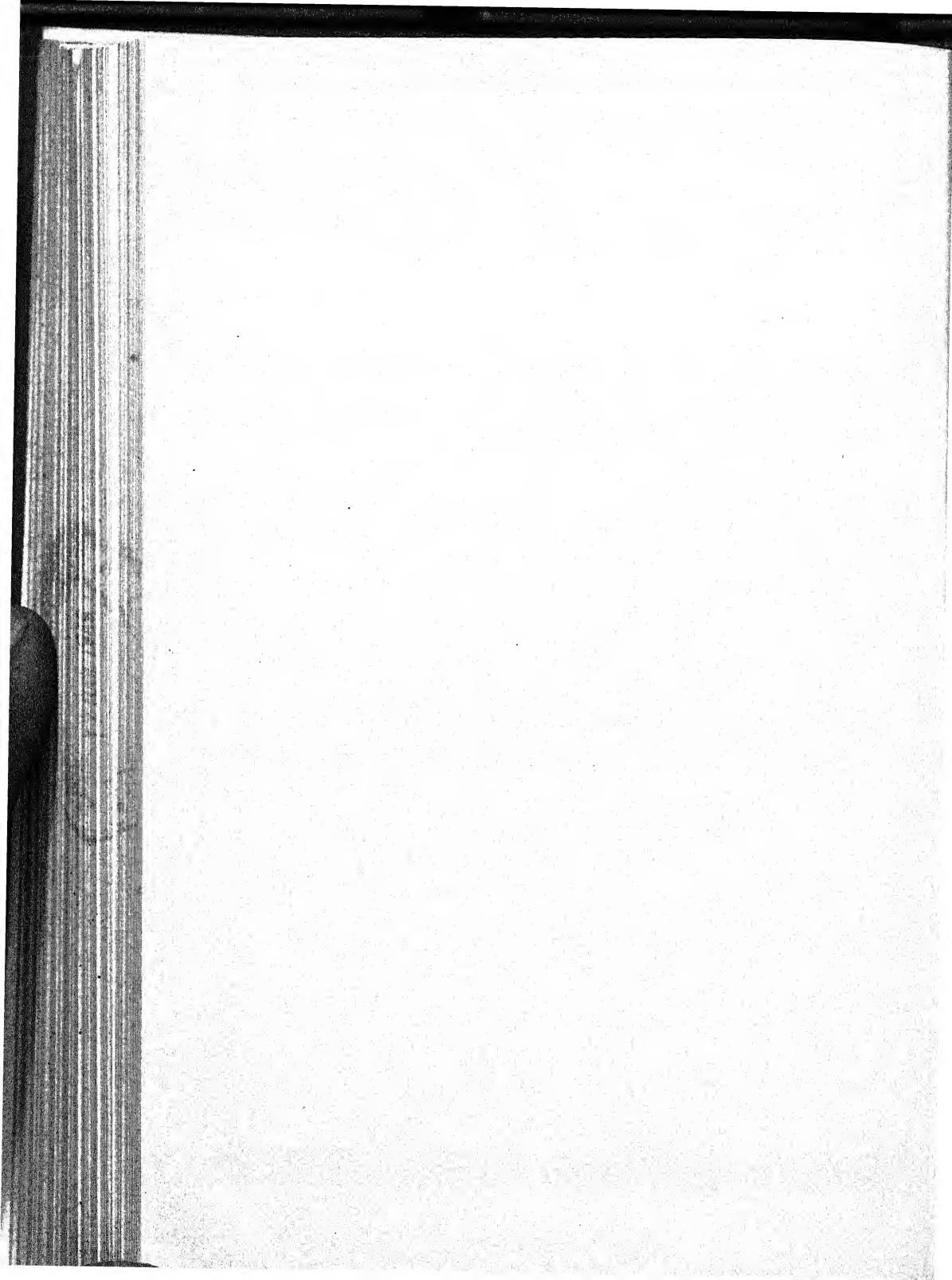
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APPENDIX

SCIENTIFIC AND COMMON NAMES OF TREES

SCIENTIFIC NAME	COMMON NAME
<i>Abies balsamea</i>	Balsam fir
<i>Acer rubrum</i>	Red maple
<i>A. saccharum</i>	Sugar maple
<i>Alnus rubra (oregana)</i>	Red alder
<i>Betula lenta</i>	Black birch
<i>B. lutea</i>	Yellow birch
<i>B. papyrifera</i>	Paper birch
<i>B. populifolia</i>	Gray birch
<i>Carya cordiformis</i>	Bitternut hickory
<i>C. glabra</i>	Pignut hickory
<i>C. ovata</i>	Shagbark hickory
<i>Celtis occidentalis</i>	Common hackberry
<i>Cercis canadensis</i>	Eastern redbud
<i>Cornus florida</i>	Flowering dogwood
<i>Fagus grandifolia</i>	American beech
<i>F. sylvatica</i>	European beech
<i>Fraxinus americana</i>	White ash
<i>F. excelsior</i>	European ash
<i>F. nigra</i>	Black ash
<i>Juglans nigra</i>	Eastern black walnut
<i>Juniperus scopulorum</i>	Rocky Mountain juniper
<i>J. virginiana</i>	Eastern red cedar
<i>Larix decidua</i>	European larch
<i>Liquidambar styraciflua</i>	American sweet gum
<i>Liriodendron tulipifera</i>	Tulip tree
<i>Magnolia acuminata</i>	Cucumber tree magnolia
<i>Nyssa sylvatica</i>	Black tupelo
<i>Picea abies</i>	Norway spruce
<i>P. engelmanni</i>	Engelmann spruce
<i>P. glauca</i>	White spruce
<i>P. rubens</i>	Red spruce
<i>Pinus banksiana</i>	Jack pine
<i>P. caribaea</i>	Slash pine
<i>P. cembroides edulis</i>	Piñon pine
<i>P. echinata</i>	Shortleaf pine
<i>P. lambertiana</i>	Sugar pine
<i>P. palustris</i>	Longleaf pine
<i>P. ponderosa</i>	Ponderosa pine

SCIENTIFIC NAME	COMMON NAME
<i>P. resinosa</i>	Norway pine
<i>P. rigida</i>	Pitch pine
<i>P. sabiniana</i>	Digger pine
<i>P. strobus</i>	Eastern white pine
<i>P. sylvestris</i>	Scotch pine
<i>P. taeda</i>	Loblolly pine
<i>Populus grandidentata</i>	Big-tooth aspen
<i>P. tremula</i>	European aspen
<i>P. tremuloides</i>	Quaking aspen
<i>Prunus serotina</i>	Black cherry
<i>Pseudotsuga taxifolia</i>	Douglas fir
<i>Quercus alba</i>	White oak
<i>Q. borealis</i>	Northern red oak
<i>Q. marilandica</i>	Blackjack oak
<i>Q. montana</i>	Chestnut oak
<i>Q. robur</i>	English oak
<i>Robinia pseudoacacia</i>	Black locust
<i>Sassafras albidum</i>	Common sassafras
<i>Sequoia sempervirens</i>	Redwood
<i>Thuja occidentalis</i>	Northern white cedar
<i>Tilia americana</i>	Basswood
<i>Tsuga canadensis</i>	Eastern hemlock
<i>Ulmus americana</i>	American elm
<i>U. thomasi</i>	Rock elm



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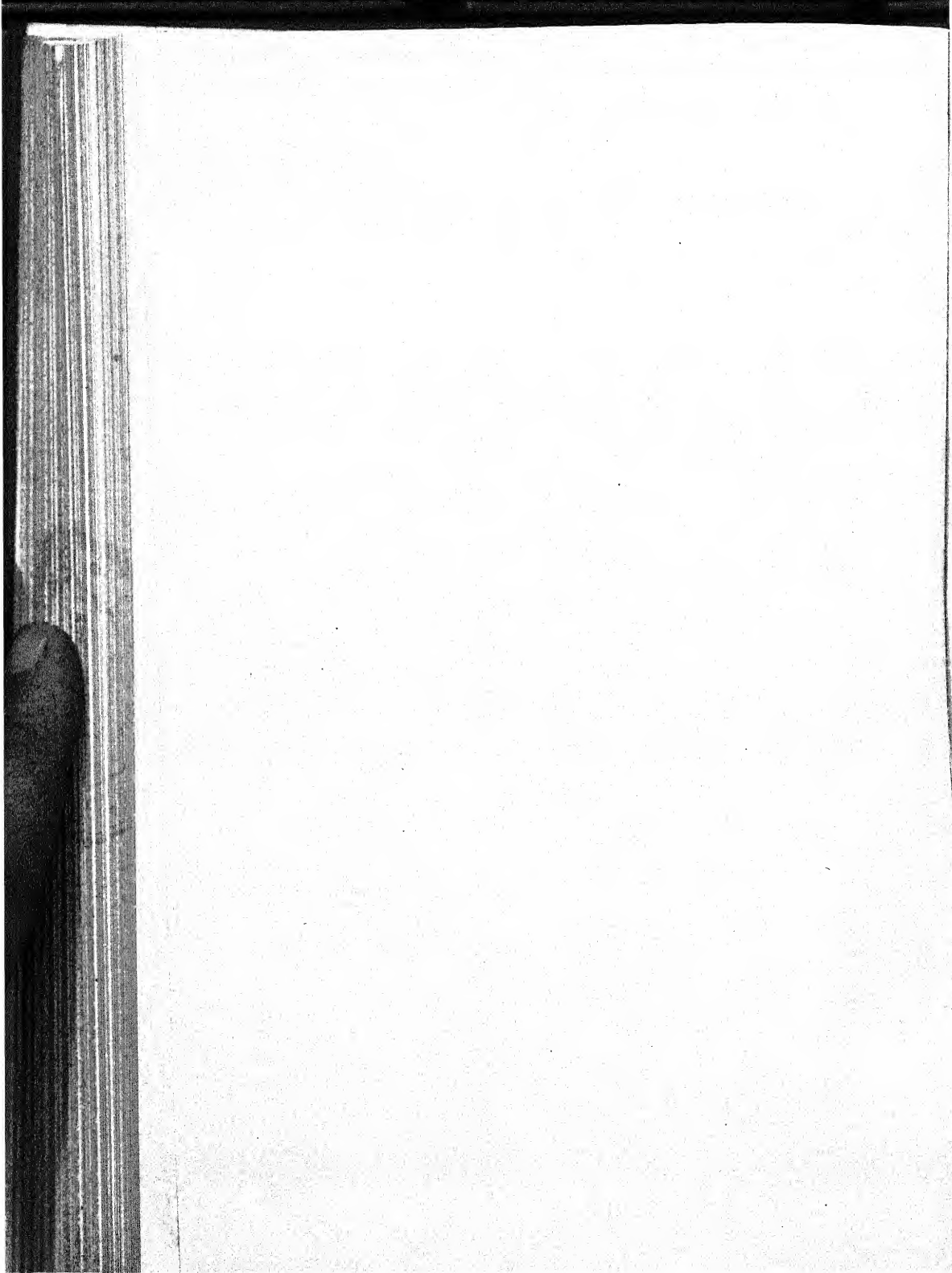
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